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STRATIGRAPHIC ZONES IN THE NEGAUNEE IRON FORMATION OF MARQUETTE COUNTY, MICHIGAN[1]

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ABSTRACT

The Negaunee iron formation, heretofore treated as a lithologic unit, may be divided into four stratigraphic zones in the area studied. Beginning with the oldest these are: the Makasin Hill zone, characterized by straight bedding throughout and abundant martite in the lower one-third; the North Lake zone, characterized by lenticular groups of strata having concretionary structure, interbedded in a straight bedded formation; the Corning Creek zone, characterized by large, lenticular, wavy-bedded bodies, interbedded in a straight bedded formation; and the Jasper Knob zone, characterized by straight bedding throughout and abundant red jasper, specularite, and magnetite or martite at least in the upper portion, and in most places extending all the way to its base. At the base of this zone there is a thick uralitic dolerite sill of which the possible genetic relation to the distinctive mineralogy of the zone is considered. Structural relations brought out by the subdivision of the Negaunee formation are also described.

OUTLINE OF PROBLEM

The general geology of the Marquette iron range in Marquette County, Michigan, has been described in several publications of the Michigan and United States Geological Surveys.[2]

[1] This article represents work done under a co-operative agreement between the Geological Survey Division of the Michigan Department of Conservation and the Michigan College of Mining and Technology, and is published by permission of the State Mining Geologist.

[2] R. C. Allen, "Revision of Huronian Correlations," *Mich. Geol. and Biol. Surv. Pub. 18, Ser. 15,* chap. ii (1915); C. R. Van Hise, W. S. Bayley, and H. L. Smyth, "The Marquette Iron-Bearing District of Michigan," *U.S. Geol. Surv. Mono. XXVIII* (1897);

PLATE I

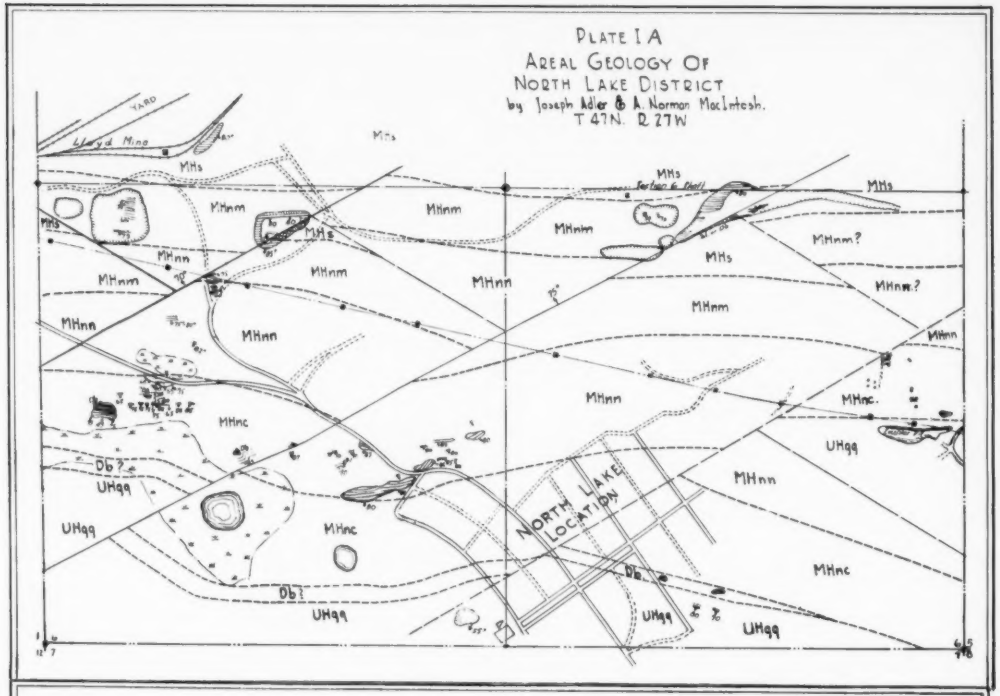
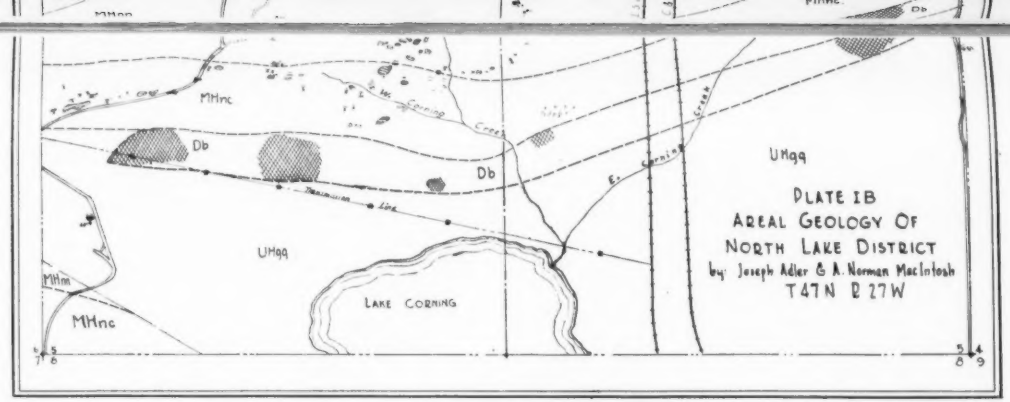


PLATE I A
AREAL GEOLOGY OF
NORTH LAKE DISTRICT
by Joseph Adler & A. Norman MacIntosh.
T 47N. R 27W
Lloyd Mine
YARD
MHs
MHnm
MHnn
MHnc
UHqq
Db?
NORTH LAKE LOCATION
MHnm?
MHnn?



DRAINAGE

- Swamps
- Streams
- Shore Lines

IGNEOUS ROCKS

UPPER HURONIAN

- Db — Diabase & Dolerite

SEDIMENTARY ROCKS

UPPER HURONIAN

- UHgq — Goodrich Quartzite

(UNCONFORMITY)

MIDDLE HURONIAN

Negaunee Iron Formation

- MHnc — Corning Creek Zone
- MHnn — North Lake Zone
- MHnm — Makasin Zone
- MHs — Siamo Slate

STRUCTURE

- 60° — Outcrops, Dip & Strike
- Contacts
- 55° — Fault showing dip
- Probable Fault
- Possible Fault

CULTURE

- Shaft
- Schools
- Main Roads
- Branch Roads
- Railroads
- Pits and Trenches
- Section Lines
- Quarter Lines
- Blueberry — Mine Names
- NORTH LAKE — Town Names
- Makasin Hills — Topography

AREAL GEOLOGY OF NORTH LAKE DISTRICT

All the workable iron deposits in the central and eastern portions of the range are included in the Negaunee iron formation of the Middle Huronian series and in the base of the Goodrich quartzite in the Upper Huronian series where the latter rests upon the Negaunee. The Negaunee formation itself is an obvious lithologic unit in places probably exceeding 1,700 feet in thickness. It is apparent that in exploring such a formation the geologist is handicapped unless he can recognize occasional stratigraphic horizons which can be used in the determination of geologic structure. The object of the research upon which this report is based was to determine whether there were any horizon markers or distinct stratigraphic zones in the iron formation.

It was immediately apparent that the ores themselves had been too greatly altered to preserve internal structures of stratigraphic significance. Phases of the ferruginous formations below ore grade may be classified under five types, namely: (1) unoxidized iron formation, (2) oxidized, but unleached, or only partly leached formation, (3) specular jaspilite, (4) jasper conglomerate, and (5) gruenerite-magnetite-schist. The unoxidized iron formation represents the least metamorphosed phase, but it was nowhere sufficiently exposed for the detailed stratigraphic studies anticipated. Since the oxidized but unleached iron formation is the next least metamorphosed type, and since considerable areas of this type are well exposed along the north limb of the Marquette Range, it was chosen as the first to receive detailed examination.

WORK DONE

Stratigraphic studies were largely confined to the belt of Negaunee iron formation outcropping in Sections 5 and 6, T. 47 N., R. 27 W., and a continuation of this belt outcropping in Sections 3 and 4, T. 47 N., R. 28 W., beyond 2 miles of swamp land. These two areas in the order named are designated as the North Lake district and the

C. R. Van Hise and C. K. Leith, "Geology of the Lake Superior Region," *U.S. Geol. Surv. Mono. LII* (1911); C. O. Swanson, "Report on the Portion of the Marquette Range Covered by the Michigan Geological Survey in 1929," Mich. Geol. Surv. Mimeographed Rept. (1930); Justin Zinn, "Report on the Portion of the Marquette Range between Humboldt and Michigamme Covered by the Michigan Geological Survey in 1930," Mich. Geol. Surv. Mimeographed Rept. (1931).

PLATE II

System	Formation		Symbol	Columnar Section	Thickness Feet	Description, Etc.
Upper Huronian	Goodrich Quartzite		UHgq		?	Coarse, vitreous, gray quartzite. Exposure incomplete.
	Diabase Sill		Db		80'	Dark greenish, coarse, chloritic, micaceous, and martitic, ferruginous to conglomeratic quartzite.
Middle Huronian	Negaunee Iron Formation	Corning Creek Zone	MHnc		470'	Wavy bedded, dense to granular, gray to pink chert with interbedded red and blue hematite and locally some martite; contains some straight bedded and concretionary portions. Generally thicker bedded than other zones.
		North Lake Zone	MHnn		660'	Chiefly straight, medium to thin bedded, or massive, red, yellow, brown and gray, dense ferruginous cherts with interbedded hematite and limonite at many horizons in the form of concretions and branching growths. Some wavy bedded portions in the upper part.
		Makasin Hill Zone	MHnm		310'	Chiefly straight, medium to thin bedded, dense chert, hematite, and limonite. Grades laterally into ore bodies. Martitic portion near base grades downward into Siamo slate.
	Siamo Slate		MHs		?	Grey to black slate with beds of gray wacke and of sideritic slate. Grades upward through hematitic slate into Negaunee iron formation. Exposure incomplete. Diabase dikes

COLUMNAR SECTION AT NORTH LAKE. SCALE 1" = 200'

PLATE III

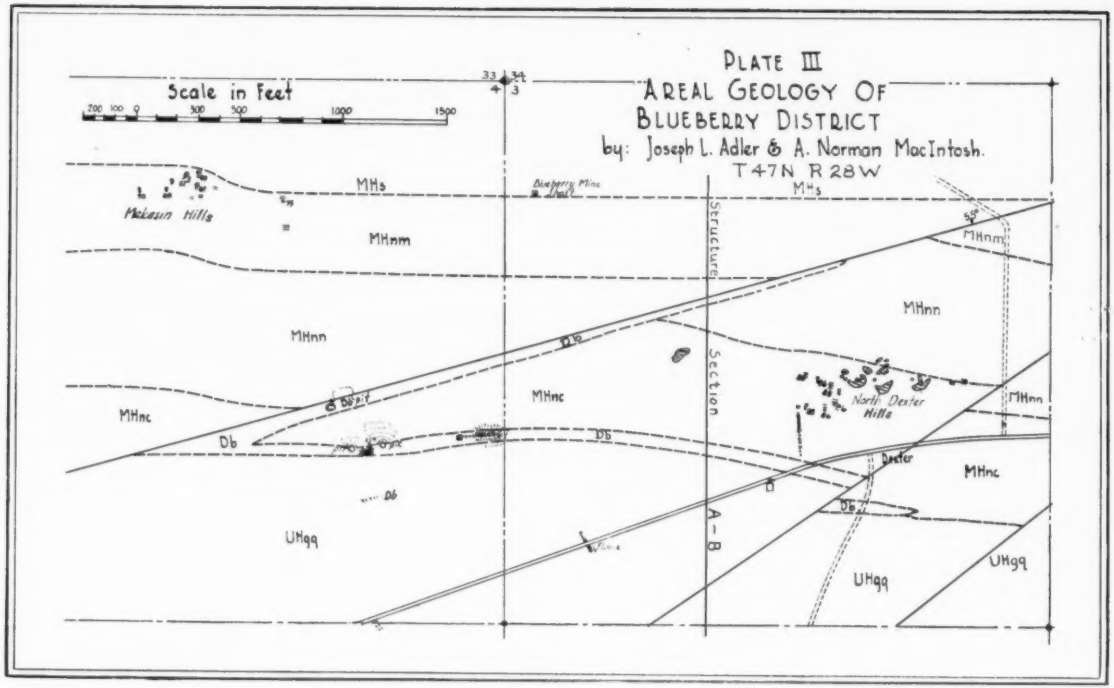
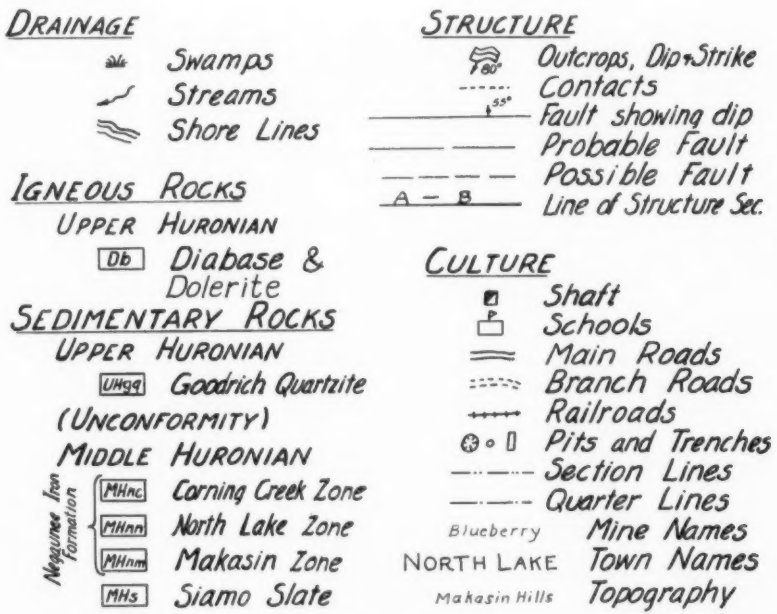


PLATE III
AREAL GEOLOGY OF
BLUEBERRY DISTRICT
by: Joseph L. Adler & A. Norman MacIntosh.
T47N R28W
Scale in Feet
200 100 0 500 1000 1500
33 34
4 3
MHs
Makasin Hills
MHnm
MHnn
MHnc
Db
UHgg
Blueberry Mine
Section
A - B
North Dexter Hills
Dexter



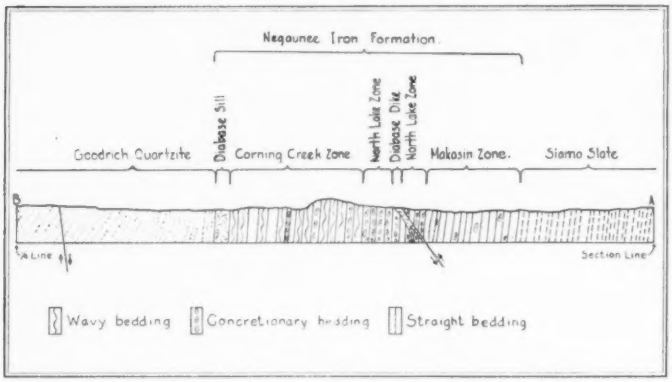
AREAL GEOLOGY OF BLUEBERRY DISTRICT

Blueberry district, respectively. A brief stratigraphic study of Jasper Knob in Sections 10 and 11, T. 47 N., R. 27 W., was also made.[3]

TEXTURES

The oxidized but unleached, or only partly leached, portion of the Negaunee iron formation is composed almost entirely of chert and iron oxides. The texture where least metamorphosed is generally too fine to permit the recognition of individual crystal grains or of dis-

PLATE IV



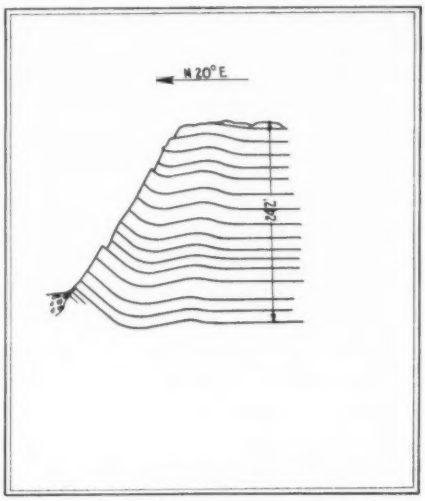
STRUCTURE SECTION ALONG LINE *AB* ON PLATE III

tinct granular aggregates with the naked eye. In the field, therefore, the fracture and luster were used to define textural types. With the exception of rare beds, concretions, and encrustations with acicular structure, the more ferruginous strata are all compact and have a smooth fracture. The chert beds, however, may be classified into

[3] ACKNOWLEDGMENT.—The author wishes to make acknowledgment of the assistance rendered by Mr. E. L. Derby, Jr., and the other officials and engineers of the Cleveland-Cliffs Iron Company; by Mr. Doell, former superintendent of the Ford Motor Company's Blueberry Mine; by Drs. W. O. Hotchkiss, C. O. Swanson, and T. M. Broderick of the Michigan College of Mining and Technology; and by Mr. F. G. Pardee of the Geological Survey Division, Michigan Department of Conservation. Special acknowledgment is due to Mr. A. N. MacIntosh for his able assistance in the field work and in the preparation of the maps.

three varieties: waxy, with a smooth fracture and waxy luster; vitreous, with a smooth to conchoidal fracture and glassy luster; and granular, with a rough or granular fracture and stony luster. The ultimate physical composition causing these differences in appearance is being investigated microscopically by Mr. A. N. MacIntosh of the Michigan College of Mining and Technology.

PLATE V



STRUCTURE NORTH SIDE OF JASPER KNOB SHOWING THICKNESS OF JASPILITE

THICKNESS OF BEDDING

The first structural character investigated was the variation in thickness of bedding. The strata were first divided into apparently homogeneous groups from 5 to 60 feet in thickness. Then the average thickness of the conspicuous bedding in each group was measured. These averages were plotted to the nearest millimeter in the form of a frequency curve (Plate VI). Inspection of the diagram shows that the curve falls naturally into two parts which have geen designated as 1 and 2. Curve Number 1 indicates that beginning with a thickness of less than 0.5 millimeter the number of occurrences increases

rapidly to a maximum for about 2.2 millimeters, then falls off only a little more slowly to a minimum for about 9.5 millimeters. Curve Number 2 surprisingly shows a considerable increase in the number of occurrences for average thicknesses of 10 millimeters and over, then a gradual decrease to zero for average thicknesses somewhat less than 40 millimeters. Curve Number 3, when extrapolated along curves Number 1 and 2 constitutes a mean frequency curve for all the points in the diagram. The point M, representing the ordinate which divides the area under curve Number 3 in half, corresponds to the median bedding thickness, if we define the median thickness as the one which is as often exceeded as not. Its value is 4.8 millimeters, which is almost exactly one-half of the thickness characterizing the break in the more closely fitted curves.

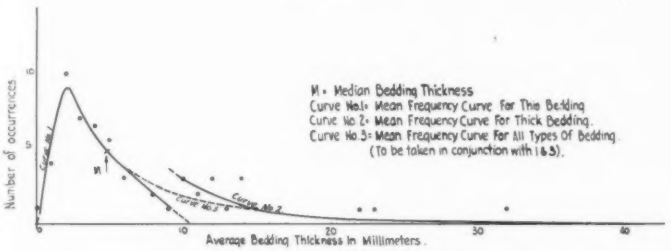
Such a sharp break in a frequency curve usually indicates that some particular factor affects the curve more pronouncedly on one side of the break than on the other. A reinspection of the groups averaging more than 9.5 millimeters in bedding thickness showed that they contained numerous discrete, conspicuous strata several centimeters thick. Strata of this type were found to be extremely rare in groups in which the bedding averaged less than 9.5 millimeters. For this reason groups of strata exceeding 9.5 millimeters in average thickness were classified as thick bedded, and groups with an average thickness of less than 9.5 millimeters were called thin bedded. Since the thin bedded groups greatly predominate, thick bedded ones, which may in many cases be traced several hundred feet laterally, are of considerable value in the correlation of adjacent drill cores, and may be used to work out local structural problems. On a regional scale, however, bedding thickness has little stratigraphic value.

In nearly all cases the iron formation exhibits also a lamination finer than the conspicuous bedding described above. The microscope reveals that it consists of paper-thin sheets of iron-oxide rich chert intercalated between more cherty layers. The more ferruginous sheets are rarely more than a millimeter apart, and their average spacing is about 0.4 millimeter. We may distinguish this finely laminated structure by the term "lamination," restricting

the term "bedding" to the broader stratification previously described.

As the bedding becomes thinner, it grows increasingly difficult to distinguish it from the lamination. No case was noted by the writer where finer laminae could be distinguished from coarser bedding in strata averaging less than one millimeter in diameter. For this reason such very thin stratification is considered identical with the lamination, and as the absence of thicker bedding gives the formation a massive aspect, it is described as "massive laminated." Finally in some rare cases, the lamination itself is too thin to be seen

PLATE VI


M = Median Bedding Thickness
Curve No.1= Mean Frequency Curve For Thin Bedding
Curve No.2= Mean Frequency Curve For Thick Bedding.
Curve No.3= Mean Frequency Curve For All Types Of Bedding.
(To be taken in conjunction with 1 & 3).
Number of occurrences
Average Bedding Thickness in Millimeters.
0
10
20
30
40

FREQUENCY CURVES SHOWING INCIDENCE OF BEDDING THICKNESS

megascopically (i.e., less than 0.1 millimeter thick) and the structure becomes simply "massive."

BEDDING STRUCTURES

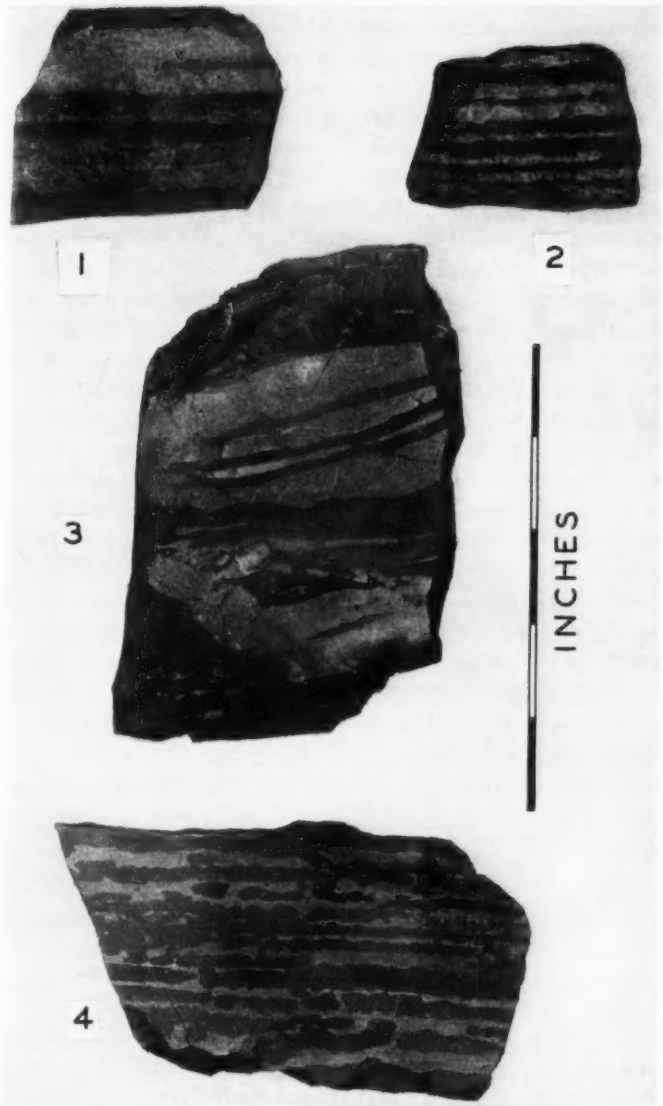
Three types of bedding were found to be fundamental, namely: straight bedding, concretionary structure, and wavy bedding. (See Plate VII.) In all three types the strata are demarcated by differences in color, luster, and hardness, due to variations in their proportions of chert and iron oxide. Straight bedding, distinguished by the parallelism of its bedding planes, is the predominant type in the Negaunee formation. In general the straight bedded iron formation is likewise thin bedded and contains waxy chert. Not uncommonly, however, it becomes thick bedded, and in a few places, particularly

where straight bedded lenses are intercalated between other types, the former contains vitreous or even granular chert.

Concretionary bedding is characterized by concretionary, bean-shaped, or branching growths of impure limonite and hematite. These growths are apparently the products of oxidation, standing out dark against a lighter background from which the unoxidized carbonate has been leached. Usually the scattered, lens-shaped concretions of iron oxide occur along definite bedding planes. In tracing such bedding planes along the strike, the observer notes first an increase in the size of the concretions, then a partial amalgamation of the edges of adjacent ones causing a type of botryoidal bedding not always distinguishable from the wavy bedding described below, and finally a complete amalgamation of all the concretions, producing a structure which approximates the straight bedding described above. In other cases the more oxidized portions deploy in branching growths into the crests of drag folds or turn off at right angles to the bedding planes along vertical joint cracks. This development may go so far that all semblance of bedding is lost with the exception of the minute laminae which, wherever distinguishable, pass through all of the described growths without deviating from their own parallel structure. Where the concretionary structure resembles bedding, the beds are mostly thin, although thick concretionary bedding also occurs. In almost all cases the chert is waxy. In the rare exceptions it is finely granular.

Wavy bedding is identified by lenses of chert, eye-shaped in cross section, which give to it its wavy appearance. The lenses are bounded above and below by thinner iron-oxide beds, which unite into single beds at each end of the lens. More rarely lenticular bodies of iron oxide are surrounded by chert. The lenses of chert and iron oxide vary from a couple of millimeters to several centimeters in thickness. The ratio of the length to the thickness of the lenses also varies greatly, with a minimum of about three to one and a maximum approaching infinity. Most wavy bedding is notably thick, although thin wavy bedding does occur. In the thinner wavy bedding the chert is inclined to be waxy, while in the more typical thick bedded type it is vitreous or granular, the thickest chert lenses being granular in nearly all cases.

PLATE VII



1 and 2, straight bedded iron formation; 3, wavy bedded iron formation; 4, concretionary iron formation.

ORIGIN OF BEDDING

The fact that the finer lamination passes undisturbed through concretions, branching iron oxide growths, botryoidal bedding surfaces, and oxidized joint walls indicates the primary character of the laminae themselves, though not necessarily the primary character of the thin sheets of iron oxide which make them visible. The origin of the so-called bedding is more doubtful. The contrast between contiguous beds evident to the eye is believed by the writer to be due to the following process. In the dark-colored beds the iron of the original ferrous carbonate was oxidized to form hematite and limonite, while the carbon dioxide was taken up in the ground water to form soluble carbonates or carbonic acid. From the light-colored beds these carbonated solutions leached the unoxidized carbonate *in toto*, the iron being removed as a ferrous compound, possibly ferrous bicarbonate. The lack of oxidation in the lighter beds may be attributed to a periodical exhaustion of available oxygen in the ground water, such as is commonly assumed to explain rhythmic banding in rocks, or to the presence in those beds of some reducing agent stronger than ferrous carbonate.

According to this explanation the difference in color is a secondary feature. But the various peculiarities of lithologic structure made evident by the color difference of the oxidized iron, such as concretionary structure and wavy bedding, are largely restricted to certain stratigraphic horizons. The author therefore believes that either some invisible sedimentary structure guides the development of these oxidized forms, or some original variations in the chemical composition of the rock predisposes certain horizons to particular types of oxidation structure.

BASIS OF SUBDIVISION

The Negaunee iron formation exhibits no evidence of intra-formational unconformities, or of stages differentiated by arillaceous or sandy deposits, such as have been used to subdivide the Middle Huronian iron formations of the Gogebic and Mesabi ranges. Variations in lithologic structure, texture, and mineralogical composition serve to distinguish the different zones recognized in this report; but in some places there are stages in which the structures characterizing adjoining zones overlap or are interbedded. For this reason it is not

always possible to identify a particular zone from a small outcrop, or from a single piece of drill core.

FOUR RECOGNIZED ZONES

For convenience the zones have been named from geographic features at their respective type localities; but as none of them has yet been traced laterally for more than 12 miles, it is not intended that the names given them be applied regionally, or even extended to the portions of the Marquette district south and west of the area studied. A detailed description of each zone follows.

Makasin Hill zone.—The Makasin Hill zone is almost entirely straight bedded. The bedding is generally thin, but on the north limb of the range toward its eastern end in the vicinity of Negaunee considerable thick bedded chert is intercalated therein. The purer chert is entirely of the waxy type, but in a large number of beds in the lower part of the zone the chert and massive iron oxides are not well differentiated. In addition to the massive hematite and limonite which form the predominating oxides of iron in all portions of the zone, abundant megascopic martite is found throughout the lower one-third. Spots may be noted which have been enriched with geodes, botryoidal beds, and veins of acicular hematite and limonite. Soft ore bodies also occur at the base of this zone.

Toward the bottom the bedding becomes thinner, so that the rock corresponds to what would be called a ferruginous slate on other ranges. On the Marquette Range the term "slate" is generally applied only to those rocks which have a sufficient clay content to develop a secondary cleavage. At the very base of the Makasin Hill zone ferruginous beds which actually have a slaty cleavage are interbedded with very thin beds of argillaceous chert lacking the cleavage. There is thus a gradation into the underlying Siamo slate.

At the type area in the Makasin Hills a very few concretionary beds and a little wavy bedding are encountered. The waviness is slight, however, and both the wavy and concretionary beds are so thin that they are nowhere conspicuous, and in the vicinity of Negaunee none are found strictly within this zone. But at the top of the zone there is a gradation into the overlying North Lake zone in which concretionary beds are characteristic. From the bottom of

the gradational beds at the top of the Makasin Hill zone to the bottom of the gradational beds at its base is approximately 300 feet.

North Lake zone.—The North Lake zone contains numerous lenticular groups of strata composed predominantly of concretionary beds, and embracing all the concretionary types previously described. The concretionary groups alternate with thicker bodies of straight bedding. This alternation of concretionary and straight bedded groups is the characteristic feature of the zone. The bedding is mostly thin. In many of the straight bedded portions it is so thin that the structure becomes massive laminated, as previously defined. On other iron ranges such beds would probably be called slate.

Individual concretionary strata have not been traced over 300 feet along the strike, although doubtful correlations were made over distances of about 800 feet. Some of the lenticular groups of concretionary beds are traceable over 1,500 feet, whereas others pinch out within much shorter distances. The individual concretionary strata and the concretionary groups, therefore, are of use in the correlation of drill cores from holes spaced at about 100-foot intervals. Though the massive laminated type of rock was found at several horizons in a drill core that penetrated the entire zone, the value of such strata as stratigraphic markers could not be determined.

The iron minerals are chiefly hematite and limonite. Hematite predominates in the straight bedded material, whereas limonite is the principal iron oxide of the concretionary growths. In the straight bedded portions the chert is characteristically waxy, but in the massive laminated type the chert and hematite are so poorly differentiated that a dull, red or grey slaty-looking rock results. The chert in the concretionary portions is waxy or rarely very finely granular. The contact with the overlying Corning Creek zone is gradational like the contact with the Makasin Hill, but is more sharply defined. Some of the concretionary lenses of the North Lake zone consists in a rapid increase in the number of wavy beds at the expense of the concretionary type, and where outcrops are numerous the contact can generally be limited to about 20 feet of strata. The total thickness of the North Lake zone measured from the bottom of its basal transitional beds to the bottom of the overlying transitional beds is from 650 to 670 feet.

Corning Creek zone.—The distinguishing structure of the Corning Creek zone is its wavy bedding. Lenticular groups of wavy bedded strata alternate with straight bedded lenses. Although thin wavy bedding is common in the zone, the conspicuously wavy bedded portions are also thick bedded, the diameter of some beds exceeding 30 centimeters. The thin wavy bedding is more easily discovered in drill cores, but it is less characteristic of the outcrops than is the thicker type. Hard, blue hematite is the predominating iron mineral. The beds of hematite are purer than those of the lower zones if we except the actual ore bodies at the base of the Makasin Hill beds. Despite the purity of the hematite beds, however, they form such a small proportion of the Corning Creek that the tenor of iron in the zone as a whole is low. Megascopically visible martite occurs at several horizons near the top of the zone. Its presence there may be attributable to the large dolerite sill which overlies it.

The chert ranges from waxy and vitreous to granular. Granular chert is characteristic of the thick wavy bedded strata, although in some cases they contain the vitreous type. The thinner bedded portions are more likely to contain waxy chert, particularly where the bedding is straightest.

The gradation from the North Lake zone upward into the base of the Corning Creek already has been described. The top of the Corning Creek is less well known. In many places it abuts against the lower side of a uralitized dolerite sill locally called "diorite." In the regions studied in detail the sill is overlain by the late Huronian Goodrich quarzite. But in the Ishpeming district to the east, the Jasper Knob zone of the Negaunee formation appears to rest upon a similar dolerite or "diorite." On account of the distinctive lithology of the iron formation overlying it, the dolerite sill has been selected as the upper limit of the Corning Creek zone. In all of the areas studied in detail, wherever the sill is missing, the Goodrich quartzite rests with unconformity directly upon the Corning Creek zone, but it is not unlikely that in much of the Ishpeming-Negaunee basin the Jasper Knob may in places rest directly upon the Corning Creek. The thickness of the Corning Creek zone, measured from the bottom of the gradational beds at its base to the bottom of the sill is about 470 feet.

Jasper Knob zone.—Owing to the resistant character of jasper, the Jasper Knob zone outcrops in many places in the Ishpeming-Negaunee basin. At the type locality in the southeastern part of Ishpeming erosion has exposed 242 feet of specular jaspilite on the north slope of Jasper Knob. That hill, indicated in Monograph XXVIII[4] as a dome, is in reality only the south half of such a structure, the north half having been cut away by erosion. (See cross section, Plate V.) According to Mr. Derby, Chief Geologist for the Cleveland-Cliffs Iron Company, only 162 feet of jaspilite were found when drilling in Jackson Grove in the southwest part of the City of Negaunee.

This zone is typically composed of thin, straight bedded layers of jasper between which are intercalated usually thinner layers of specularite. The non-specular iron oxide is chiefly martite and magnetite. Where the writer examined outcrops of this zone, only a few lenses of very subordinate extent contain white or grey chert and blue hematite, as opposed to red jasper and specularite, but on Mr. Derby's evidence, cited in the foregoing, the writer believes that more such material may exist in the lower portion of the zone around Negaunee. A few soft-ore bodies occur at the bottom of the zone, resting on the dolerite sill.

Because of its uniformly straight bedded structure and its distinctive mineralogy in most places, the Jasper Knob zone cannot generally be confused with the other zones of the Negaunee formation in their normal condition, but it somewhat resembles the "jasper conglomerate," so commonly associated with the Goodrich-Negaunee contact. The jaspilite of the Jasper Knob zone is well bedded, however, whereas the jasper conglomerate is composed of heterogeneously oriented fragments of specular jaspilite in a matrix of specularite, chert, and sand. The latter is usually considered the end product of the dynamo-metamorphism of a hematitic or limonitic residual soil developed by weathering in the pre-Goodrich erosion interval. It has been derived by these processes from many different zones in the Negaunee formation.

As in the case of the conglomerate, the prevalence of specularite and jasper in the Jasper Knob zone is probably attributable to the

[4] Van Hise, Bayley, and Smyth, *op. cit.*, Atlas, Pl. 28.

presence of hematite or limonite in the rock in place of siderite before the final dynamo-metamorphism of all the Huronian formations. But the more definite stratigraphic position of the Jasper Knob zone and its distinct lithologic structure indicate that these ferric oxides came into existence under conditions different from those which oxidized the conglomerate.

Three theories accounting for the origin of these ferric oxides are advanced. The first and simplest propounds that the iron was originally precipitated from seas or lakes in the form of limonite or hematite, as in the case of most Paleozoic and later iron ores. The second theory, propounded by Van Hise and Leith,[5] supposes that during the epoch preceding the pre-Goodrich dissection of the Middle Huronian strata, a low water table permitted the weathering of the Negaunee formation *in situ* to proceed to a considerable depth. The third theory is directly connected with the intrusion of the dolerite sill. It is known that the contact metamorphism produced by sills and laccoliths is largely confined to overlying rocks because of the tendency of the heated magmatic waters and the volatile solutes contained therein to rise. It is, therefore, possible that magmatic waters ascending from the sill oxidized most of the overlying iron formation, particularly where the flow of these waters was impeded by the base of the less permeable Goodrich quartzite. The oxidized iron formation could then either concomitantly or subsequently be altered to specular jaspilite by dynamo-metamorphism. This theory involves a slight modification of the one lately suggested by Gruner.[6] Because of the limited amount of data thus far collected on the Jasper Knob zone, no choice will be ventured of the three theories here presented to explain its distinctive mineralogy.

In the two places where the base of the Jasper Knob zone was observed it lay upon the dolerite sill. Except where denuded by recent erosion, the zone is everywhere unconformably overlain by the Goodrich quartzite, the contact generally being marked by a body of hard ore in the form of conglomerate. Since the Jasper Knob zone lies entirely outside of the area which this report was originally intended to cover, its geology has been studied only superficially.

[5] *Op. cit.*

[6] J. W. Gruner, "Hydrothermal Oxidation Experiments," *Econ. Geol.*, Vol. XXV (1930), pp. 840–44.

PRACTICAL APPLICATIONS OF THE STRATIGRAPHIC DATA

In conclusion there will be presented some of the structural features which the writer was enabled to determine by recognition of the different stratigraphic zones in the Negaunee formation. On both limbs of the range the Negaunee formation is cut by two systems of faults, mostly normal.[7] Some of these faults sufficiently offset the Negaunee formation as a whole so that their existence may be deduced from this criterion alone, but in most cases the piecing together of outcrops of the individual zones is needed to reveal the faulted relationship. In the North Lake area (Plates IA and IB) an older fault which dips northeasterly about 35° is offset by two or three faults of the younger system. The fault intersections produce several triangular patches of the Makasin Hill and North Lake zones. Since the former rests on a footwall of Siamo slate, the overlapping of the slate where the faults cut its upper surface forms plunging troughs of the type that control soft-ore concentrations, some pitching steeply to the southwest, others more gently to the southeast. Leaching of the Makasin Hill zone in these troughs has produced the known soft-ore bodies of the North Lake district. It should be noted that the probable faults indicated to the east of the area already mined might form similar ore-bearing troughs.

In the Blueberry district (Plate III) there are at least three faults of the younger system. The northernmost is a reverse fault along the upper surface of a diabase dike. The intersection of this dike with the upper surface of the Siamo slate would form a trough inclined gently to the west which may account for the Blueberry ore body. Attention is called to the fact that the other faults found in the Blueberry district may produce additional troughs to the east of the area mined. Likewise the northward shift in the surface of the slate as it is traced westward beneath the Makasin Hills forms another trough.

The writer believes that a thorough study of available drill cores would reveal a number of new marker horizons within the different zones here recognized. Such markers would greatly facilitate interpretation of the geologic structure. During the investigation more faults developing ore-bearing troughs on top of the Siamo slate might be located.

[7] Swanson, *op. cit.*

RELATIONS OF ANORTHOSITE TO GRANITE

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University of Minnesota

ABSTRACT

A comparison of the anorthosites of the Adirondacks with those of Minnesota indicates that the difference lies largely in the intense deformation of the Adirondack rocks, during or after crystallization, by which the relations of rock are obscured. At Duluth, and probably in most anorthosite districts, the granitic phases of magma seem to have evolved from gabbro magma some time after the anorthosite was evolved. Both are related to gabbro but not differentiated directly from each other. Gradations from granite to anorthosite originate by inclusion and reaction.

INTRODUCTION

Few of the excursions of the Sixteenth International Geological Congress were announced as having any special interest for petrologists; but wherever buses stopped at rock outcrops, the chorus of pick and hammer indicated that petrologists were at work on their problems. Two of the excursions, both well attended, led to discussions that may be worthy of a note. The anorthosites of the Adirondacks and those at Duluth were visited by men familiar with such rocks in Canada, Europe, Asia, and Africa. These two districts in the United States are very different; but at each the foreign visitors found familiar features—there is little but what is characteristic of anorthosites in other countries. The Bushveld mass most nearly resembles that at Duluth, and several Canadian and European districts resemble the Adirondacks.

The chief contrast between the two occurrences is probably structural, for the Duluth mass suffered little deformation after crystallization was well advanced, whereas the whole Adirondack region has been so deformed that scarcely a rock is free from granulation and minerals suggesting metamorphism. Even some pegmatites, though probably not all, in the anorthosite are sheared and deformed. This deformation is a common feature of anorthosites elsewhere,[1] but fortunately a few masses like that at Duluth have escaped.

[1] N. L. Bowen, "The Problem of the Anorthosites," *Jour. Geol.*, Vol. XXV (1917), pp. 218–19.

One of the puzzling features of such rocks is the occurrence of large gray labradorite grains (crystals, fragments, or augen) in the whole series of rocks that seem to be genetically related—anorthosite gabbro, syenite, and granite. These include the syenite of the Adirondacks and the "red-rock" granite of Duluth. The problem seemed so obscure in the eastern mass, and the answer so convincing to the whole group of visitors at Duluth, that the writers were stimulated to add some details and present a few comments.

PETROGRAPHY AT DULUTH

The relation of rocks at Duluth is described in the Guidebook of Excursion C 4,[2] as follows, with Plate 6 to show the contrast in textures.

> At Thirteenth Avenue West a small abandoned quarry shows some very instructive relations between the red granite and the anorthosite. On casual inspection it seems that there is as complete a gradation from red rock to anorthosite here, as there is from red rock to gabbro in Lincoln Park. Close study shows, however, that the anorthosite solidified earlier than the red rock, and that the late red rock phase of the magma shattered and probably assimilated some anorthosite. There is no true gradation from one to the other.

The visitors agreed with the writer of the guide that the gray feldspars did not grow in the magma after it had evolved to a state capable of forming red rock, but at a much earlier stage when, if chilled, the rest-magma would have been of basaltic composition. They agreed that the anorthosite exhibited a strong tendency to shatter[3] down to xenocrysts that spread out in progressively smaller amounts in the red rock farther from the solid anorthosite source. They agreed that if dynamic action had been superposed on the rocks seen, the evidence of the history would have been so obscured that the two end-members of the series would seem to have been differentiated directly from each other instead of having been mixed after differentiation.

The extent and completeness of the series should be emphasized. Analyses of red rock and anorthosite have long been available. Table I includes also tests of two of the contaminated rocks at Thirteenth Avenue West. The several constituents suggest that

[2] *XVI Int. Geol. Cong. Guidebook 27* (1933), p. 71.

[3] This is true in other districts where anorthosite is intruded by magmas.

Sample 3 contains somewhere from 40 per cent to 85 per cent of anorthosite (average 61 per cent). Without doubt some of the rock in the quarry is even more largely anorthosite than the 85 per cent indicated by the silica content of Sample 3. If such a rock is a contaminated red rock, the contaminating xenocrysts made up several

TABLE I

ANALYSES OF ROCK SPECIMENS FROM DULUTH, MINNESOTA

Constituents	(1) Red Rock	(2) Red Rock Slightly Contaminated by Anorthosite	(3) Red Rock Much Contaminated by Anorthosite	(4) Anorthosite
SiO_2	66.92	.....	51.96	49.39
Al_2O_3	12.51	.....	18.97	29.08
Fe_2O_3	4.36	.....	1.98	0.34
FeO	3.93	.....	7.07	2.89
MgO	1.66	.....	2.67	2.26
CaO	1.20	.....	8.60	13.06
Na_2O	3.45	3.93	3.18	2.89
K_2O	3.98	3.40	1.43	0.10
H_2O+	1.25	.....	1.31	0.34
H_2O-	0.20	.....	0.17	0.09
TiO_2	0.69	.....	1.90	trace
P_2O_5	0.11	.....	0.41	0.09
MnO	0.16	.....	0.14	0.04
Others	0.34	.....	.....	.....
	100.76	.....	99.79	100.57
	II·4·2·3	.....	II·5·4·4	I·5·4·5

1. Analysis of specimen of red rock selected from the type locality. F. F. Grout, analyst; *Jour. Geol.*, Vol. XXVI, No. 8 (1918), p. 650, No. 25.

2. Alkalies of red rock slightly contaminated by anorthosite. W. W. Longley, analyst.

3. Analysis of red rock much contaminated by anorthosite. W. W. Longley, analyst.

4. Analysis of anorthosite. F. F. Grout, analyst; *Jour. Geol*, Vol. XXVI, No. 8 (1918), p. 646, No. 3.

times the volume of the magma in which they were suspended. See Figure 1.

Careful work was also done on the indices of the feldspars (Fig. 2). The anorthosite feldspar is about $Ab_{40}An_{60}$, and the larger gray xenocrysts in the red rock have the same composition. Thin sections show that there has been a great deal of reaction, making the outer irregular zones more sodic, some having two or more rather

distinct zones. The most sodic material found in large gray xenoliths by the use of index oils was $Ab_{45}An_{55}$, but the thin sections show more sodic zones—possibly with a change of color. The extinction angles of the outer zones may be 20° different from those of the core.

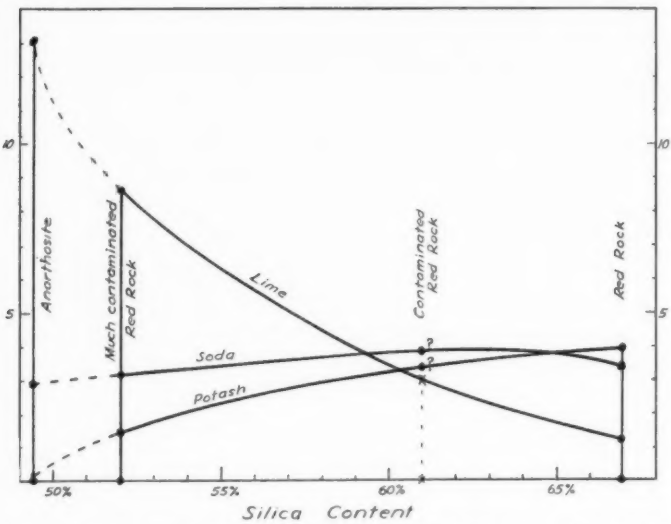

FIG. 1.—Curves on a basis of silica content, showing the lime, potash, and soda of red rock, anorthosite, and red rock contaminated with anorthosite.

On the other hand, the red feldspar of the red rock is progressively affected by the xenocrysts. The fresh red rock has much orthoclase, $\alpha = 1.527$, $\gamma = 1.531$, in graphic intergrowth with quartz, and a little red plagioclase with maximum symmetrical extinction about 15° and indices, $\alpha = 1.532$, $\gamma = 1.539$. As xenocrysts become numerous the red matrix shows less color. Close to the anorthosite only a few pinkish gray feldspars can be detected. Very little graphic intergrowth is found in rocks with 50 per cent or more xenocrysts. The pink feldspar of highly contaminated rock has indices from 1.526 to 1.540, being perhaps, oligoclase with some potash in it.

Several thin sections showing xenocrysts of labradorite in red rock reveal an additional event in the history of the feldspars. Not

only are the calcic feldspars attacked by reaction with the red-rock magma and changed, to form less calcic outer zones; but when the red feldspars began to grow, at least a part grew on the xenocrysts as secondary enlargements. This is the more certain because the outer layer is graphically intergrown with quartz, and clearly characteristic of red rock. It is not a myrmekite, or a deuteric attack of quartz on the outer zone of the xenocryst.

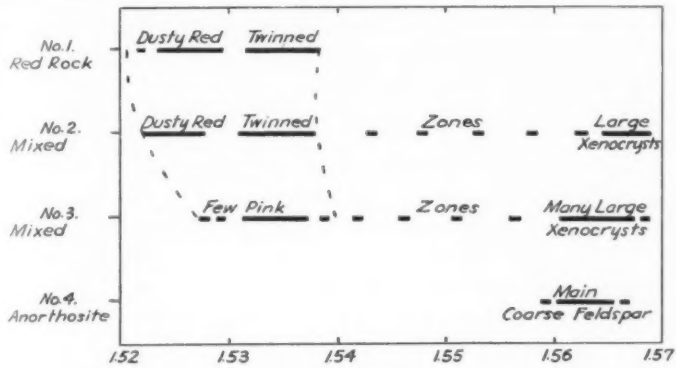

FIG. 2.—Diagram of the indices of feldspars in red rock, anorthosite, and some red rock that is contaminated or mixed with anorthosite.

INFERENCES FROM DULUTH

It seems clear from the field agreement and the recent laboratory work that the basic plagioclase here accumulated to form a solid rock before the red-rock magma came to place. When the magma of red rock was brought in contact with the anorthosite (probably still hot), it shattered the anorthosite and began active reaction with the fragments. This could hardly have been prolonged over such a period of time as was involved in the original growth of basic plagioclase to coarse crystals. It is perfectly clear, therefore, that the basic plagioclase did not grow in a magma of the composition of red-rock magma. The rest-magma of anorthosite was almost certainly a large volume of gabbroic nature, which later differentiated by a somewhat different process to form a rest-magma of small volume and granitic composition. If the early crystals of the anorthosite had a rest-magma of more sodic feldspar in equilibrium with a

more calcic feldspar, they reacted to approximate uniformity as they finally solidified.

A similar history was worked out at Pigeon Point,[4] where the early plagioclases formed not only small anorthosite masses but phenocrysts in a groundmass of diabase. The plagioclases of phenocrysts and groundmass were both labradorite, showing that even in a sill of moderate size the feldspar phenocrysts that might accumulate as anorthosite were in reaction equilibrium with the same feldspar in the groundmass. The more silicic red-rock feldspars formed later from a part of the magma that remained liquid, while the ferromagnesian minerals probably settled.

APPLICATIONS

This sequence which seems so clear in Minnesota has not been advocated by any of the several geologists working in the Adirondacks. That area is a more difficult one, because the mass is greatly deformed, and its base is not well exposed. Nevertheless, it may be said that there is not any strong evidence against such an idea as that suggested at Duluth, in any of the Adirondack reports, or in any outcrops visited by the Congress excursion.

Labradorites in large crystals are found in three or more very different rocks in the Adirondacks, namely, in the anorthosite, in the gabbro, and in the syenite—all now deformed into augen so that they seem to have much the same relations to their surroundings. It hardly seems likely, however, that labradorite phenocrysts could grow in two magmas as different as syenite and gabbro. It is much more likely that fragments already crystallized could be carried in a syenite magma (without complete reaction to more sodic forms) to almost any distance.[5]

Bowen,[6] in a most valuable paper emphasizing the probable origin of anorthosites by the accumulation of crystals, offers a suggestion as to the association of granite, anorthosite, and gabbro, which re-

[4] F. F. Grout, "Anorthosite and Granite as Differentiates of a Diabase Sill," *Geol. Soc. Amer. Bull.* 39 (1928), pp. 555–78.

[5] Balk (p. 384) in the work cited later, notes that they occur 7 miles from an anorthosite outcrop, and Buddington writes that fragments may be found scores of miles from the main anorthosite.

[6] Bowen, *op. cit.*, pp. 209–43.

lates the syenite about as closely to anorthosite as to gabbro. This depends on a nice adjustment of specific gravities, such that labradorite crystals grown in a magma remain suspended until the magma changes its composition (without much changing the labradorite). The labradorite then forms a solid by the collection of crystals, leaving siliceous magma behind. Two field evidences raise questions as to this outline: first, there are many diabases in which early labradorite crystals rise rather than remain suspended;[7] and second, there are commonly signs that the anorthosite was solid rock before the disturbances brought syenite or granite magma in contact with it. At Duluth both dikes of red rock and xenoliths of anorthosite are common, but in the Adirondacks the dikes are so sheared that they have been questioned.

Balk's paper[8] is the largest one of recent years on the Adirondacks, and is a masterpiece of structural information and interpretation. He calls attention (pp. 340 and 396) to the gap in composition of feldspars in large and small plagioclases in the syenite, noting that there is a similar gap between crystals and liquid in laboratory crystallization of plagioclases. "The rocks of the syenite series are considered the frozen mother liquor of the magma." This wholly ignores the fact that in such a large mass the early crystals should react with the magma to at least approximate uniformity. Balk notes that any possible fragments of anorthosite ought to react with a more silicic rest-magma (p. 395), but seems to overlook the fact that a hot crystal growing in such a magma should have more chance to react than a fragment. Furthermore, if the labradorite grew in a syenite matrix, how could the same labradorite have grown in a gabbro matrix and an anorthosite matrix? It seems much more probable that the gap observed is a result of fragments of one sort being included in a magma of another sort, without time for complete reaction.

Barth[9] and Buddington[10] seem to take about the same attitude as

[7] Grout, *op. cit.*, p. 565.

[8] Robert Balk, "Structural Geology of the Adirondack Anorthosite," *T. Min. Pet. Mit.*, Vol. XLI (1931), pp. 308–434.

[9] T. F. W. Barth, "Adirondack Feldspar," *Amer. Mineralogist*, Vol. XV (1930), pp. 129–43.

[10] A. F. Buddington, "The Adirondack Magmatic Stem," *Jour. Geol.*, Vol. XXXIX (1931), pp. 240–63.

Balk in regard to the rest-magma of the labradorite crystals which make up the anorthosite, although Buddington would hardly think that the syenite magma came out from the anorthosite, because the evidence of intrusion and inclusion is too clear. The syenite is a late phase injecting an earlier phase anorthosite.

There is no disposition anywhere to question the physical chemistry which shows that labradorite crystals are in equilibrium with a more sodic feldspar melt, or mother-liquor; but the field evidence rather suggests that the calcic plagioclase which formed anorthosite accumulated with a small amount of more sodic matrix almost free from dark minerals, and then, as it finally solidified, reacted with that matrix forming the uniform feldspar now found. At that stage the main magma from which the anorthosite separated was probably still large in relative volume and still almost gabbroic in composition. The separation of basic crystals from this might leave a syenite magma to inject the anorthosite.

Alling's discussion[11] of the problem is most recent and comes perhaps the closest to our scheme of petrogeny. He finds that in the Adirondacks the nordmarkite (syenite) minerals have replaced the anorthosite minerals, by reaction.

Balk assumes in many statements that anorthosite and syenite have separated from each other,[12] and that the syenite and pegmatite have been squeezed out of the anorthosite rather than injected into it. He presents a long argument against the idea of dikes of the more silicic rocks cutting anorthosite, saying "not true dikes," "not intrusive in the strict sense" (pp. 376–77), and "pseudo intrusive" (p. 389), and later calling them "veins." This is just a matter of terms. He agrees that they solidified later than anorthosite and may have moved after the anorthosite was more nearly solid, but he seems to fear that if they are called dikes it will imply a distinct supply of magma rather than a related phase of the same magma. The history here suggested does not imply a different magma source,

[11] H. L. Alling, "The Adirondack Anorthosite and Its Problems," *Jour. Geol.*, Vol. XL (1932), pp. 193–237, especially pp. 214–23. Alling's doubt as to the age relations of the gabbro is also explained in our scheme, for some gabbro may be cooled from primary magma before anorthosite accumulated and some differentiated from the syenite-granite later.

[12] *Op. cit.*, pp. 357, 372, 373.

but neither does it admit a direct separation of anorthosite from syenite. The observed gradation from one rock to the other may be almost complete in either case, but the origin of the series is notably different—in the one case by squeezing out, in the other by introduction of a late phase into an early solidified one, with shattering and reaction.[13]

Such rocks as are undeformed igneous products in a mass like that at Duluth should not be attributed to deformation if they are found also with similar associates in the Adirondacks. It would be wise for those who attempt the study of petrology in so complex a region as the Adirondacks to study first the simpler districts. In such districts both anorthosite and the syenite-granite series are genetically related to gabbros, but anorthosite formed early and was associated with granite only after a further differentiation of the gabbro magma to a siliceous rest-magma.

[13] Balk objects to the idea of xenocrysts because the labradorites in large gray crystals in syenite are not corroded. This needs further observations. He says mechanical forces can hardly succeed in isolating phenocrysts from anorthosite, but the fact is that *they do*. The field evidence is clear in Minnesota that anorthosite has a conspicuous tendency to shatter down to xenocrysts, whether the intruding magma is red rock or diabase; and this is no doubt just as true in other parts of the world.

THE GLACIAL HISTORY OF AN EXTINCT VOLCANO, CRATER LAKE NATIONAL PARK[1]

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Clark University

ABSTRACT

Alternating layers of lava and glacial material in the rim surrounding Crater Lake record the history of an intermittently active volcano on the slopes of which glaciers formed during periods of dormancy, to be destroyed when igneous activity resumed. The position of buried moraines and the pumice character of material mantling the glacial topography indicate that Mount Mazama disappeared as the result of collapse. The glacio-volcanic sequence discovered at Crater Lake is particularly significant in that it throws light upon the history of the volcanoes of the Northwest which may be similar to that of Mount Mazama before its destruction.

THE GLACIO-VOLCANIC SEQUENCE

Hidden away in the volcanic rocks of the Cascade Range of southern Oregon is the record of Mount Mazama, an ancient volcanic cone that grew to great height and later disappeared entirely, leaving a giant caldera in which the deep-blue waters of Crater Lake have since accumulated (Fig. 7). The story of this mysterious mountain is recorded in the rocks of the region. Like leaves in a book, the alternating layers of lava and glacial material in the rim surrounding Crater Lake tell the story of the late monarch of the Cascade Range.

During the vulcanism of mid-Tertiary time small volcanic cones developed in the Cascade region, one of which was destined to become Mount Mazama (Fig. 1). With continued igneous activity the youthful mountain attained sufficient altitude to cause heavy precipitation on its slopes. Snows accumulated and remained through succeeding seasons. Glaciers were born and the ice fields moved slowly down the slopes of the intermittently active volcano (Fig. 2). Evidence of these early glaciers is found in the form of till deposits

[1] Published by permission of the director of the National Park Service.

The author is indebted to Clark University for financial assistance which made possible the completion of the Crater Lake studies started in 1931 under the auspices of the National Park Service.

buried beneath several hundred feet of volcanic material and younger glacial débris.

Glaciers on the slopes of a volcano sooner or later are apt to fall victim to renewed lava eruption. In the case of young Mount

FIG. 1.—The youthful Mount Mazama as it may have looked during the early stages of its growth. Continued volcanic activity gradually produced a mountain.

FIG. 2.—A later stage in the growth of the volcano. The cone is dormant and small glaciers are present. Successive stages of vulcanism and glaciation followed.

Mazama, the glaciers were destroyed several times during the growth of the mountain. The glacial landscape of Figure 2 was changed to the volcanic landscape of Figure 3. Gradually this activity subsided and the scene reverted to a glacial landscape (Fig. 4). This succession of changes may be called the glacio-volcanic sequence.

While the pen-and-ink sketches show only one sequence, the glacial deposits exposed in the rim surrounding the present lake indicate that four, and possibly many more, stages of glaciation were inter-

FIG. 3.—Mount Mazama during one of its last periods of volcanic activity. A secondary cone, Little Mazama, is situated on the western slope.

FIG. 4.—The last glacial landscape. The U-shaped valleys which notch the present rim were produced during this final ice invasion. The dotted line indicates the location of the rim.

spersed with the periods of vulcanism. This continued glacio-volcanic sequence is suggested in Figure 6.

It will be noted from the sketches that the mountain at first had but one cone, while during the later stages of volcano-building a secondary cone developed. This smaller mountain is Little Mazama.

The absence of this cone during the glacial stage illustrated by Figure 2 and its existence during later stages in the growth of the mountain are determined by the direction of striae markings found

FIG. 5.—The present Crater Lake located in the giant caldera, formed by the collapse and engulfment of Mount Mazama. The Wizard Island cinder cone developed after the disappearance of the mountain.

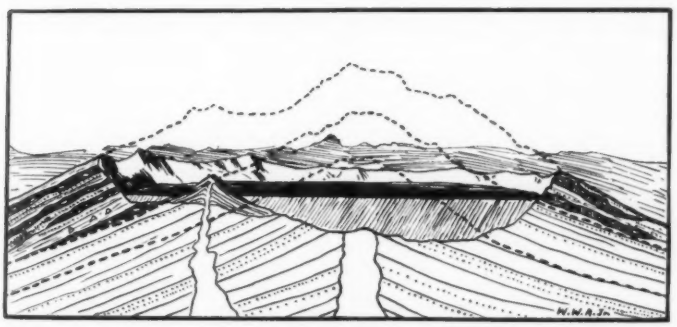

FIG. 6.—A generalized cross section of the region today. The alternating layers of till and volcanic material record the story of the growth of Mount Mazama. The dotted lines mark the several stages represented in the preceding drawings.

buried at different levels beneath lavas and glacial débris (see Figs. 23 and 24).

Without discussing at this place each of the glacio-volcanic sequences, we arrive at the stage in the growth of the mountain represented by Figure 4. The volcano was apparently dormant, and large

glaciers radiated from the summits of Mount Mazama and Little Mazama. These glaciers started a rapid destruction of the cone. Volcanic activity began again in a shower of pumice which whitened the entire landscape. Many of the high points on the rim, as well as the morainic deposits left by the last glaciers, are still partially buried beneath this pumice. But in spite of the enormous amounts of pumice scattered over the landscape, the disappearance of the entire mountain mass above the dotted line in Figure 4, and much of the core to a depth of 3,000–4,000 feet below the present rim, cannot be explained by explosive eruption alone.

Approximately 15 cubic miles of material have disappeared in order to produce the landscape shown in Figure 5. If explosion accomplished this great change, we would expect to find a thick mantle of fragmental andesitic lava and breccia scattered over the surrounding countryside. Furthermore, the glacial deposits, formed by the glaciers which existed only as long as the mountain existed, should be heavily buried with the fragmental material derived from the destruction of the mountain. Instead we find very little angular material in the region, and most of that had its origin prior to the last glaciers which existed on the mountain before its destruction.

Unlike Krakatoa and Katmai, Mazama did not blow itself to pieces; but instead, it is believed, this mountain collapsed and was engulfed. As early as 1901 Josef S. Diller proposed this theory in his presidential address delivered before the Geological Society of Washington.[2] A year later the results of Diller's field work appeared as a professional paper of the United States Geological Survey.[3] Although it is difficult to conceive of such a phenomenon, field evidence to date affords no acceptable alternative. The mountain certainly existed; the mountain is now gone, and the 15 cubic miles of material have not been found. The processes of engulfment were probably slow. They may have been similar to the caving-in which takes place in Hawaii where the huge calderas are from time to time being enlarged by engulfment.

[2] J. S. Diller, "The Wreck of Mount Mazama," abstract published in *Science*, N.S., Vol. XV (January–June, 1902), pp. 203–11.

[3] J. S. Diller and H. B. Patton. "The Geology and Petrography of Crater Lake National Park," *U.S. Geol. Surv. Prof. Paper 3.*

FIG. 7.—*Above*, Wizard Island and the western rim. The massive flow producing the high cliff beyond and to the left of Wizard Island is Llao Rock.

Below, the rugged eastern rim with large talus slopes reaching the water's edge. Phantom Ship appears in the foreground.

Following the formation of the giant caldera near the close of the Pleistocene, or shortly thereafter, there was a brief period of inactivity interrupted by the building of Wizard Island cinder cone and two other smaller cones reported to exist on the floor of the present lake. Since the completion of the Wizard Island Cone, there has been no volcanic activity in the immediate region.

FIG. 8.—Looking southeast from a point above and to the west of the Devils Backbone. The remarkable U-shaped valley of Sun Creek, visible in the photograph, is the result of glacial scour immediately preceding the collapse of Mount Mazama. (Official photograph, U.S. Army Air Corps.)

With the cessation of vulcanism a lake formed in the bottom of the caldera. The annual precipitation far exceeded the amount of water lost each year by evaporation and seepage, and consequently the lake level rose. Crater Lake is now nearly 2,000 feet deep, and it maintains a relatively constant level throughout the year. A certain amount of water disappears through underground channels and reappears in numerous springs on the lower slopes of the mountain base.

IMPORTANT LOCALITIES

In order to unravel the long and complicated history of Mount Mazama, many field observations were made on the precipitous

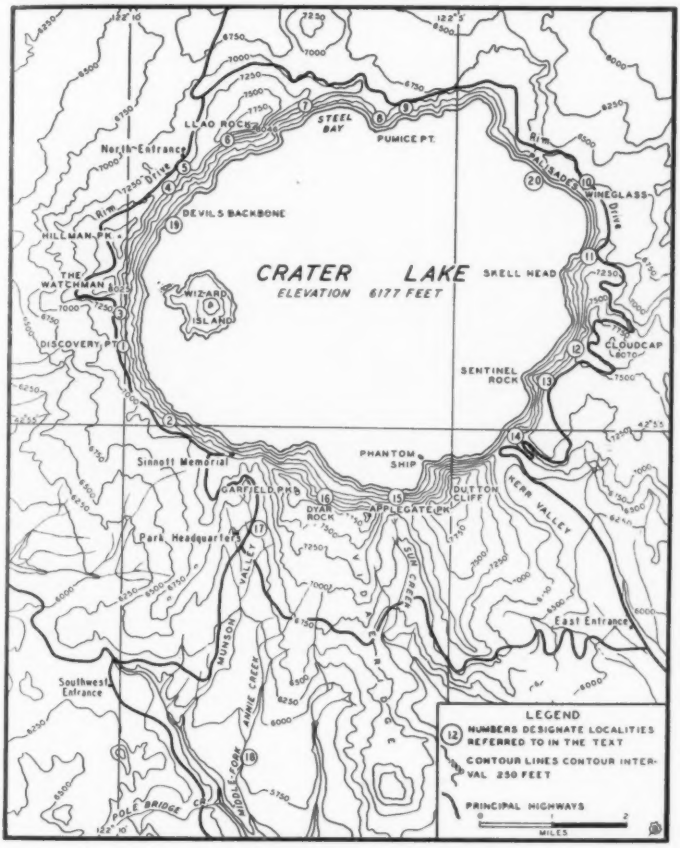

FIG. 9.—Index map of the Crater Lake region

cliffs surrounding the lake. A few of the significant localities will be described and related to the glacio-volcanic sequence which preceded the collapse and engulfment. The locations of all exposures mentioned are shown on the reference map, Figure 9.

Discovery Point (1).—On the rim at Discovery Point, partially covered by overlying pumice material, are several beautifully polished and striated rock surfaces (Fig. 10).[4] The existence of these striae on the surface lavas indicates that glacial ice crossed this portion of the rim following the last lava eruption.

When standing on the striated surface close to the edge of the cliff, it is possible to look down approximately 40 feet to another

FIG. 10.—Glacially striated surfaces at the edge of the rim just west of Discovery Point.

glacially polished surface. Upon investigation it is discovered that these glacial scratches are covered with 10 feet of boulder till and that the till is, in turn, capped with 30 feet of lava. This relationship is best shown in Figure 11. How did the glacial material become sandwiched in between flows of lava? The answer is wrapped up in the growth of Mount Mazama. The lower glacial horizon is related to a stage of glaciation which was abruptly terminated when the mountain erupted the lavas which now cap the till deposits. The upper glacial horizon dates to the last glacial stage which oc-

[4] All photographs, unless otherwise credited, are official pictures of the National Park Service.

curred after the surface lavas had cooled and before the mountain collapsed.

FIG. 11.—The type locality at Discovery Point. The striated surface of Figure 10 occurs on the rim a few feet to the right of this photograph. The boulder till interbedded with the lavas indicates a stage of glaciation which was terminated by volcanic eruption.

Glacier Point (2).—Close to the trail which follows the rim east of Discovery Point is an excellent exposure illustrating three stages of glaciation. Three lava flows of different ages exhibit striae (Fig. 12). The lower surface is capped with till containing subangular and striated stones, while the other levels are relatively free of glacial débris.

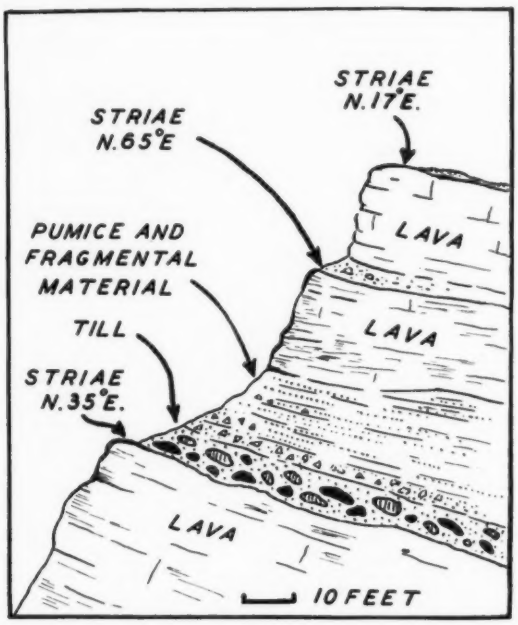

FIG. 12.—Three stages of glaciation at Glacier Point. Locality 2

South of the Watchman (3).—About half way between Discovery Point and the Watchman, resting upon a striated lava surface, is a thin layer of glacial till and in turn some 50 feet of stratified pumice and fragmental material. In the pumice is a carbonized log standing in upright position. The stump and roots appear to have decomposed, allowing a portion of the log to settle. Mr. D. S. Libbey, the Park naturalist, in collaboration with Albert E. Long, excavated the

log during the summer of 1933. It is believed that the tree was growing in a thin layer of glacial till when volcanic ash and pumice buried it. The roots and base of the tree were buried first by cool pumice, but subsequently hot volcanic ash and pumice settled around the tree so fast that air was excluded, combustion was prevented, and carbonization resulted[5] (Fig. 13).

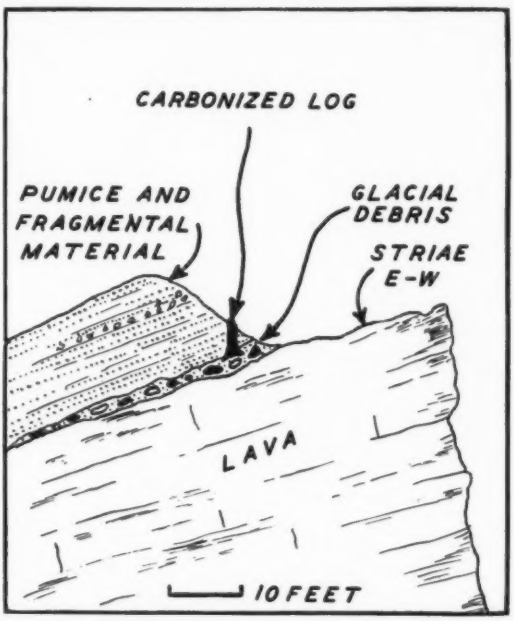

FIG. 13.—A carbonized log buried beneath pumice and fragmental material. Locality 3.

North of the Devils Backbone (4).—Beside the rim road a short distance north of the Devils Backbone is a beautifully polished and striated surface of lava (Fig. 14). Resting upon this is a thin layer of till and a deposit of pumice, locally several feet thick and interbedded with angular fragments of andesitic lava. In places the pumice rests directly upon the well-striated surface.

[5] D. S. Libbey, "Carbonized Tree Found within the Rim," *Nature Notes, Crater Lake National Park*, Vol. VI, No. III (August, 1933).

FIG. 14.—Glacial striae on the lavas just north of the Devils Backbone. The surface is highly polished and, except where the rock has chipped, owing to weathering, the striae are very distinct. The ice moved in a N. 17° W. direction across this lava surface.

FIG. 15.—A much weathered, glacially grooved surface a short distance south of the North Entrance Ranger Station. Although the age of the grooves is difficult to determine, they appear to be related to a stage of glaciation which ended with the return of volcanic action on the mountain.

Near North Entrance Ranger Station (5).—Beneath pumice and what appears to be a much fractured lava formation is an old weathered surface which exhibits well-developed glacial grooves (Fig. 15). The poor preservation of the surface and its buried location suggest that the markings are older than those found just north of the Devils Backbone. If talus material did not cover so much of the inner slope at this point, it would probably be possible to relate this

FIG. 16.—Llao Rock viewed from the south. Beneath the massive dacite flow at a point marked *A* is an exposure of glacial till. At *B* a similar deposit rests upon a striated surface. The contact between the lava and underlying till suggests that ice existed on the mountain during the first stages of the eruption which produced Llao Rock.

grooving to the earlier of the glacial stages recognized at Discovery Point.

Llao Rock (6).—From a point on the rim just east of the North Entrance Ranger Station it is possible to view the steep front of Llao Rock. If one is equipped with a pair of good boots, it is possible to crawl along at the base of the steepest portion and reach a position marked *A* in Figure 16. Here, buried beneath the lava of Llao Rock, is a heavy boulder till containing numerous well-striated stones ranging from a few inches to a foot in diameter. The material is characteristically glacial in appearance. A few hundred feet farther

down the slope, and several hundred yards to the south, is a glacially polished bedrock surface overlain by boulder moraine. The position of these various evidences of glaciation beneath the dacite flow which produced Llao Rock is conclusive proof of glaciation on Mount Mazama prior to the eruption which produced the Llao Rock flow. From a vantage point on the lake or on the opposite rim, an excellent view of Llao Rock is obtained which shows very clearly the topography which existed before the Llao Rock flow descended the slopes of Mazama (Fig. 17). On the basis of the glacial evidence reported

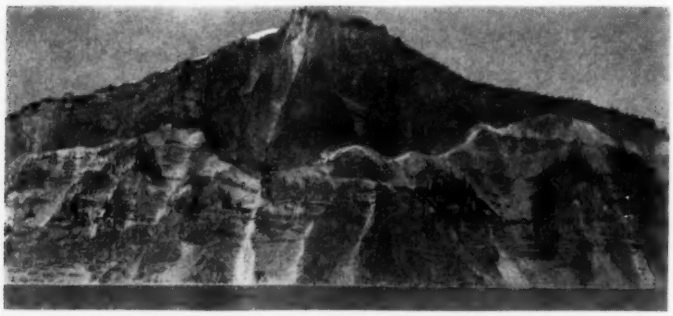

FIG. 17.—The massive Llao flow which descended upon a glacial landscape filling the U-shaped valleys and burying the morainic débris. The lower portion of the cliff which rises almost vertically from the water's edge is composed of alternating layers of lava and volcanic ejectamenta recording the growth of the young Mount Mazama.

above, the pre-Llao topography presented a glacial landscape and the U-shaped form of the central portion of the Llao flow is due to glacial scour in the valley, which was later occupied by the lava stream. Because of inaccessibility it was impossible to determine whether glacial material exists beneath the central portion of the flow.

Steel Bay (7).—After considerable difficulty a section was investigated at the western edge of Steel Bay. The very steep walls prevented complete examination, but the record uncovered was nevertheless of value. There are two distinctly striated surfaces—one buried beneath approximately 100 feet of pumice and the second several hundred feet down the cliff.

Pumice Point (8).—Across Steel Bay to the east is Pumice Point, easily identified from almost any position on the rim because of its large white pumice face. On closer inspection, however, it is discovered that there is more than pumice on Pumice Point (Fig. 18). Interbedded in the volcanic material are layers of glacial till and old soil. The sequence of events has evidently been one of glaciation,

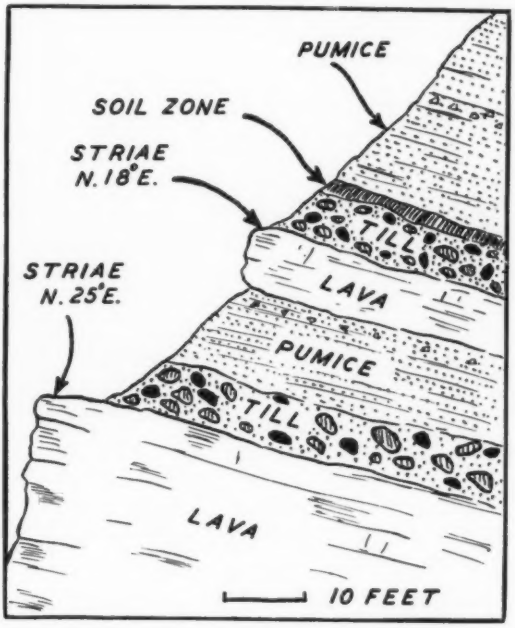

FIG. 18.—Two glacial stages at Pumice Point. A soil zone appears in the upper portion of the younger till layer. Locality 8.

vulcanism, and renewed glaciation, followed by an interval during which soil developed and then by another period of vulcanism.

The upper portion of the younger till layer is a dark-colored soil zone containing an abundance of vegetal material, much of it charred. The existence of the soil layer on the steep pumice talus slope was first suggested by a band of vegetation which appeared on an otherwise barren hillside. By removing the veneer of pumice, the

moisture-retentive soil zone was discovered. The relationships here are similar to those on the rim south of the Watchman, although the carbonized material at Pumice Point is limited to small fragments. Future excavations may uncover larger and better-preserved representatives which, it is hoped, can be identified. The real significance of the soil zone in the story of Mount Mazama is the time interval implied. Like the periods of vulcanism and glaciation, the periods when soil developed and vegetation became established on the mountain required time. These records make it possible to reconstruct partially the time intervals in the growth of Mount Mazama. The lava surfaces exposed beneath the glacial till were well striated in a N. 32° E. direction.

Pumice Drive (9).—A short distance northeast of Pumice Point and 75 feet below the level of the highway is a layer of till 5 feet or more in thickness, containing many well-striated stones, some of them as much as 2 feet in diameter. The lava beneath the glacial material is so weathered that all record of striae has been long since destroyed. Above the till are 30 feet of stratified pumice followed by 10–15 feet of lava breccia and 30 feet of pumice. The conditions here suggest that considerable eruption followed the last glacial advance in this locality.

The Wineglass (10).—North of Cloudcap and east of the Palisades is an interesting feature known as the Wineglass. Viewed from the lake, the white talus material appears like a huge goblet, the constricted portion at the base of the bowl being produced by a resistant layer of columnar lava. Resting on this columnar lava is a deposit of glacial till consisting of a variety of volcanic rocks, many of them subangular and striated. Many of the larger stones are 2 feet in diameter. Above the till are approximately 20 feet of pumice followed by 10–15 feet of lava breccia and 25 feet of pumice. Bedrock striae were not observed at this locality.

Skell Head (11).—A short distance from the rim road northeast of Skell Head a thin deposit of glacial till rests upon striated bedrock. The till contains nicely rounded and striated stones, some of which suggest the work of water as well as of ice. Compass readings indicate that the ice moved in a N. 50° E. direction across the bedrock surface.

Cloudcap (12).—On the rim directly west of Cloudcap is a layer of pumice and fragmental material well over 100 feet in thickness resting upon a striated lava surface. Owing to the inaccessibility of the bedrock exposures and the advanced stage of weathering, only two striae readings were recorded, both indicating a N. 70° W. direction. Unlike most localities thus far discussed, no layer of distinctly glacial material was found. Numerous subangular stones suggest that at least a portion of the fragmental material overlying the glacial surface was transported by ice.

Sentinel Rock (13).—A mile to the southwest of Cloudcap is Sentinel Rock, a promontory readily recognized from almost all outlook points on the rim. Here a resistant lava formation is buried under layers of pumice and fragmental material. In 1931 excellent striae were located on a partially loosened block close to the edge of the cliff. By allowing for the amount of displacement, a reading of N. 75° W. was obtained. In 1933 the same locality was revisited, but the block had disappeared. In hopes of locating new exposures, search was made to the south along the contact between the lava and overlying fragmental material. While bedrock striae were not found, owing to the weathered character of the exposed rock, an excellent boulder moraine was discovered. The lower layer of fragmental material which was resting upon the lava at Sentinel Rock gave way to glacial till containing nicely polished and striated stones. Above the till was a thick layer of stratified pumice and occasional bands of coarser material. Even with absence of bedrock striae, evidence of a buried glaciation was convincing.

Kerr Notch (14).—On the steep cliff of Danger Bay, just north of Kerr Notch and 500 feet below the glacial material referred to above is a second buried glacial deposit. Striated stones ranging from small pebbles to boulders 2 feet in diameter are common. The upper portion of the till is roughly stratified, suggesting that rains or glacial flooding accompanied the deposition. Above the glacial material is a layer of columnar lava followed by alternating layers of pumice and breccia; below it is a much weathered lava on which no striae were found.

In addition to buried evidence of glaciation the U-shaped profile

of Kerr Valley offers proof of ice action (Fig. 19). Like the glaciated surfaces west of Cloudcap, the U-shaped floor of Kerr Valley is buried with stratified pumice and fragmental material. Since the stratification appears undisturbed, it is to be inferred that glaciers did not exist in these areas during or after the eruption of pumice. Judging from the heavy accumulation of pumice at Cloudcap, as compared with the 30–40 feet at Kerr Notch, glacial ice probably remained in the valley for some time after it had abandoned the higher land to the north.

FIG. 19.—Crater Lake viewed from the northwest. At the left is Mount Scott, the highest elevation in the Park. On the far side of the rim is the U-shaped valley of Sand Creek, usually referred to as Kerr Notch. (Official photograph, U.S. Army Air Corps.)

Sun Notch (15).—Between Applegate Peak and Dutton Cliff is the broad U-shaped valley of Sun Creek (Fig. 8). Below the usual layer of pumice is a striated lava surface over which the ice rode. At several locations between the highway and the rim, morainic features were recognized, although most of them were more or less effectively buried by pumice and fragmental material. As at Kerr Notch, the relationships suggest that the glacier abandoned the valley prior to the last pumice eruptions.

Dyar Rock (16).—A short distance west of Sun Notch in the col between Garfield and Applegate peaks, a layer of glacial till rests upon a striated lava surface. Compass readings indicate that the ice proceeded in a N. 10°–20° E. direction. Above the till is a layer of pumice and fragmental material similar to that found in Sun

Notch to the east. Back from the rim where the till and pumice covering has been partially removed by wind and rain, some bedrock outcrops exhibit striae. One and a half miles south of the rim, close to the highway, good morainic topography indicates that glacial ice once covered part, if not all, of the slope west of Vidae Ridge.

Munson Valley (17).—Directly east of the government camp and superintendent's office in Munson Valley is one of the best morainic evidences of glaciation to be found in the Park. While striae are

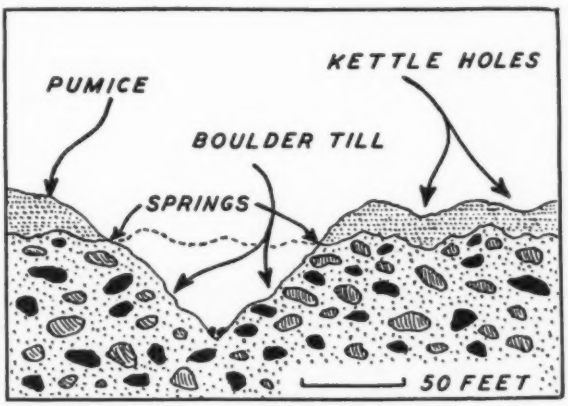

FIG. 20.—Buried kame and kettle topography on the Middle Fork of Annie Creek. Locality 18.

very scarce, the curving form of the ridges and their hummocky topography leave no doubt as to their glacial origin. Unlike most of the glacial features thus far reported, these moraines are not buried by pumice.

Middle Fork Annie Creek (18).—Three miles south of the government camp between Middle Fork and Annie Creek is a most interesting topography resembling kame-and-kettle moraine. Directly opposite, on the other side of Annie Creek along the highway north of Pole Bridge Creek, is a similar topography. In both areas there are numerous hills of pumice separated by depressions in pumice. Not until a cross section of the material was discovered in the valley of Middle Fork a little over a mile above the junction of that stream with Annie Creek did an explanation come to light (Fig. 20). In the

bottom of the V-shaped gorge were hundreds of large polished boulders that certainly did not originiate from the pumice formation. On closer examination several of these boulders were found to be striated. With this clue a careful study was made of the walls of the gorge. The upper 20–30 feet were composed of dry pumice, while the lower 40–50 feet were soaking wet and grass-covered. Numerous springs issued at the top of the grassy portion, making it extremely difficult to negotiate the steep slope. The line of springs marks the top of relatively impervious morainic material from which the boulders in the stream bottom were derived.

The walls of this gorge provide the cross section of the kame-and-kettle topography. With the underlying material identified as boulder till, it is safe to assume that the puzzling hills and depressions resulted from glacial deposition and that the pumice veneer is but a postglacial camouflage.

The full significance of this buried hummocky moraine may rest in its suggestion of the terminal position of the Munson Valley glacier. To be sure, there is no proof that the deposit does not represent a recessional stage of the glacier; and no evidence thus far obtained would deny a much more extensive glaciation than is here described. Some day the valley of Annie Creek may reveal new clues; but until definite evidence of ice action is found farther down the valley, there is no reason for assuming that ice proceeded beyond the position indicated on Middle Fork. The possibility of securing new evidence at greater distances from the rim is very slight in view of the vast pumice and ash deposits which bury so much of the former glacial landscape.

South of the Devils Backbone (19).—Within the rim just south of the Devils Backbone and only 300 feet above the lake level is the oldest record of glaciation discovered in the region (Fig. 21). Buried beneath 800–900 feet of lava and volcanic material of various kinds is a deposit resembling till, although striae do not appear. Underlying this formation, however, is a striated bedrock surface indicating very definitely that glaciers existed on Mount Mazama quite early in its growth. One naturally wonders how high the mountain

was during this early stage, but adequate evidence on which to base any calculation is entirely absent.

Palisades (20).—A short distance to the west of the Wineglass is a massive flow which produces a precipitous cliff 300–400 feet high, known as the Palisades. At the foot of this cliff is a long talus slope underlain in part at least by heavy boulder moraine (Fig. 22). Approaching the exposure from the lake, the subangular form of the

FIG. 21.—A portion of the rim showing the large black dike known as the Devils Backbone. The location of the oldest glacial record is indicated by the lower striae reading. The striae on the rim are those pictured in Figure 10 and were produced by the last glaciers to descend the west slope of Mount Mazama.

boulders is the first characteristic to be noted. Closer inspection reveals striae. The surface on which the till rests is undoubtedly beneath the level of the lake. This glacial deposit is lower in elevation than any found within the rim, but it is not so deeply buried under volcanic material as the old striated surface south of the Devils Backbone.

PROJECTED STRIAE

In plotting the striae readings found on the rim, certain very interesting facts came to light. All striae that appeared to be associ-

ated with the last advance of ice were recorded on the map. This selection included the striae that were buried by pumice and fragmental material, but not those beneath lava flows. When the readings were complete, the lines indicating direction were projected toward the lake, with the result shown in Figure 23. It is not safe to place too much significance on this type of mapping; nevertheless, certain results appear convincing. Most of the lines projected from

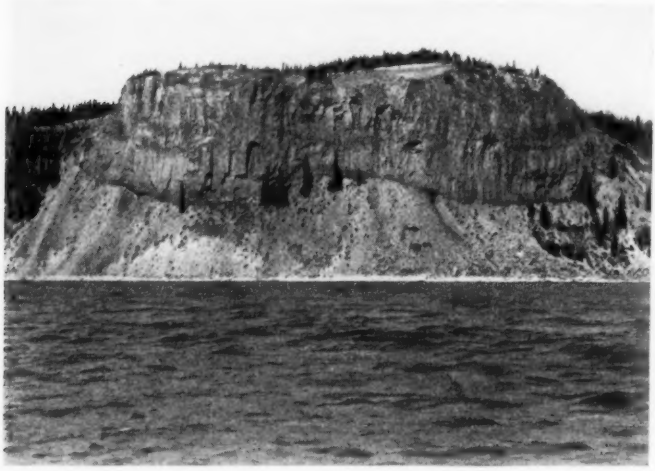

FIG. 22.—The Palisades as seen from the lake. Beneath the talus material is a glacial till containing many well striated boulders. The bottom of the deposit is presumably below the level of the lake.

readings on the north, south, and east rims roughly converge above the center of the lake. This fact, coupled with the volcanic evidence in the rocks surrounding the lake, suggests rather definitely that the summit of the original Mount Mazama was somewhere above the center of the present Crater Lake.

A second center of intersection of projected striae appears over Wizard Island. This is certainly no mere coincidence but instead seems to prove the existence of a secondary cone which, like Shastina no the side of Mount Shasta, grew on the slope of Mount Mazama. The ice which crossed the rim between Discovery Point and the

Devils Backbone undoubtedly originated on this peak which once existed above the present Wizard Island (Figs. 4 and 5). If this is the case, Wizard Island is probably the result of renewed activity from the same channel from which, in an earlier period, came the lava which produced Little Mazama.

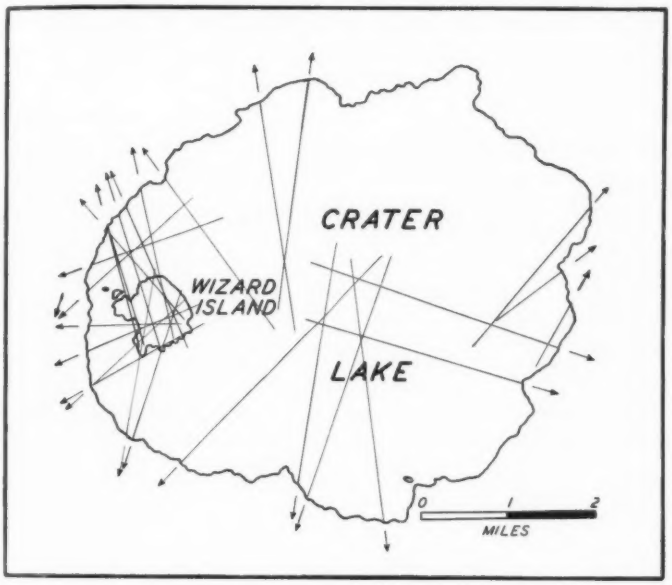

FIG. 23.—The direction of ice movement during the final stages of glaciation on Mount Mazama. The projected striae suggest that there were two centers of ice dispersal—one somewhere above the middle of Crater Lake, and the other above Wizard Island.

Following this method of investigation a little further, another map has been constructed showing the direction of the striae which exist within the crater rim and are definitely related to the earlier stages of glaciation (Fig. 24). For example, the reading recorded at Pumice Point is 300 feet below the rim, and the one just south of the Devils Backbone is over 800 feet below the rim or 300 feet from the water's edge. While the number of striae readings are necessarily fewer than those available for more recent glacial stages, the results

obtained from their projection may be of equally great interest. The projected striae roughly converge above the center of the lake, again suggesting a central position for the main peak of the mountain of Mount Mazama down whose slopes the early glaciers moved. No projected striae converge over Wizard Island. This indicates that

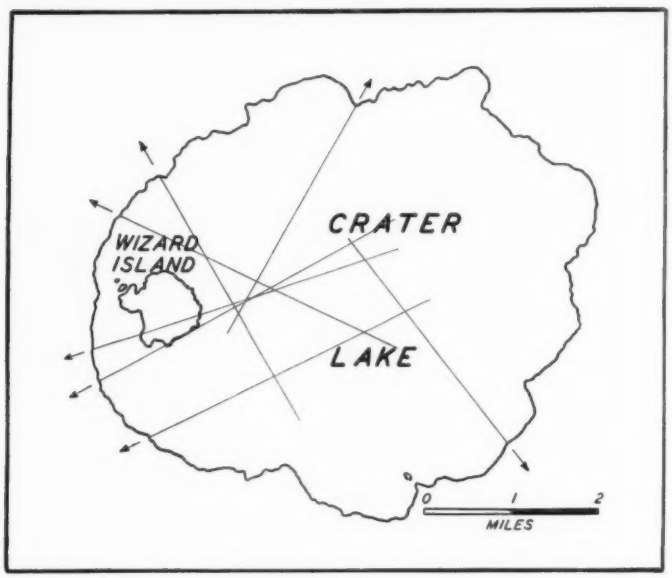

FIG. 24.—The direction of ice movement during the early stages of glaciation on Mount Mazama. The projected striae suggest that the ice spread outward from a mountain, the central peak of which was somewhere above the middle of the present lake.

no secondary cone had come into existence when the early glaciers were present. Any more definite conclusions must await further study of the volcanic history of the region.

THE SIGNIFICANCE

A glacio-volcanic sequence similar to that so clearly recorded in the rocks surrounding Crater Lake has probably occurred in the history of Mount Ranier, Mount Hood, Mount Shasta, and other

volcanoes of the Cascade Range; but the evidence is not visible. Although Mount Ranier is dormant at present, the glaciers on its slopes are always in danger of being destroyed by a renewed volcanic eruption. Thus, previous generations of glaciers may have been destroyed. If the summit of Mount Ranier were suddenly to disappear, leaving a giant caldera in the base of the mountain, many of the conditions observed today at Crater Lake would very probably be duplicated. The collapse of Mount Mazama opened a book for the geomorphologist which might otherwise have remained closed forever.

From the standpoint of the glaciologist and vulcanologist the results of the Crater Lake study are not limited in application to the Cascade Mountain region but apply to all volcanic areas where the summits rise to sufficient altitude to allow glaciers to form or where previous conditions have favored glaciation. Assuming that the stages in the growth of Mount Mazama preceding its collapse were normal for high volcanoes during the Pleistocene, we should expect to find similar glacio-volcanic sequences in many of the mountain regions throughout the world. Unfortunately, the records of other volcanoes are so completely hidden that a sequence comparable to that discovered at Crater Lake may never be available for study. Evidence of glacio-volcanic sequences may, however, be uncovered in the walls of the stream valleys which radiate from large volcanoes. It is very probable that the valleys which dissect the lower slopes of Mount Ranier hold records of this kind.

In addition to the light which Mazama's story may cast upon the history of other volcanoes, there is a still greater significance. Each glacial deposit exposed in the rim indicates that snow accumulated on the mountain and that ice fields developed of sufficient size to descend the slopes, polishing and deeply striating the rocks. The time required to accomplish this task is necessarily great, probably many thousand years. The four known glacial records interbedded with volcanic materials indicate four long intervals in the growth of Mount Mazama.

The development of a soil horizon like that at Pumice Point required much time. Each layer of volcanic breccia overlain by a lava flow implies that the volcano changed from an explosive type

to a quiet eruptive type. The occurrence of dacite instead of andesite during the late stages of mountain growth signifies internal changes. Each successive layer of volcanic material exposed in the rim required time to accumulate, and each one that was followed by glaciation required time to cool before the process of snow accumulation could start,

Thus the glacial record of an extinct volcano helps to establish time intervals in volcano-building and contributes to the knowledge of earth history.

THE "CLINTON" IN WESTERN NEW YORK

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ABSTRACT

A study of the stratigraphy and paleogeography of a portion of the "Clinton" formations occurring between the Niagara Gorge and Martville, New York, points to the following conclusions. The Reynales, including the hematite beds, the Martville sandstone, and other members, is a single formation, although it is neither a time nor a lithologic unit. The Neahga, Lower Green shale of Hall, at Niagara, and the Maplewood shale of Chadwick at Rochester are of limited extent along the strike, indicating possible independence. The purple shale at Rochester and Sodus Bay should not be divided but constitutes a single unit, the Sodus formation of Chadwick.

INTRODUCTION

The Clinton of New York State includes sediments believed to have been deposited in the shoreward regions of oscillating seas, and hence contains frequent lithologic, faunal, and stratigraphic changes. These irregularities in the character of the deposits have resulted in several varying interpretations since the area was first studied in any real detail by James Hall[1] in 1843. The resulting formulations of the stratigraphy can best be shown by tables (Tables I, II, and III).[2]

The present paper is limited to the area between the Niagara Gorge and Martville, and does not attempt to discuss the entire Clinton, even in this restricted area. A study of the sections to the east of Martville is still incomplete, and the upper portion of some of the sections to the west can best be discussed in relation to them. The problem of the limits of the Clinton, some suggestions regarding which were made by me[3] at a recent meeting of the Paleontological Society, likewise awaits discussion until a more thorough study has been made of the eastern sections.

The relationships of the Neahga, not certainly known except in the Niagara Gorge, are considered, as are those of the Maplewood

[1] James Hall, *Geology of New York, Part 4, Survey of the Fourth Geological District* (1843).

[2] These tables include several formations above and below the Clinton.

[3] Paper presented before the Paleontological Society, December, 1932.

shale, which is not definitely recognized in the outcrops away from the vicinity of Rochester. These two formations occupy the same approximate horizon and may be of similar age. The Reynales is traced across the area and is shown to be a stratigraphic unit, although its fauna, lithology, and age differ from place to place. At Rochester a purple shale is found above the Reynales. This is shown to belong to the same stratigraphic unit as the Sodus shale in the vicinity of Sodus Bay.

I am indebted to friends at the University of Rochester and at Princeton University for much help and advice, especially to Professors Harold L. Alling and J. Edward Hoffmeister of the former, and to Professors B. F. Howell and R. M. Field of the latter. Assistance was also given by the Buffalo Museum of Science.

THE NEAHGA FORMATION

At the Niagara Gorge the lower Clinton green shale is 6 feet thick; it thins eastward and is nearly, if not completely, missing at Lockport.[4] It was originally correlated by Hall[5] with the green shale below the Pentamerous limestone at Rochester (Maplewood shale), to which it corresponds in stratigraphic position. These shales with the Sodus of Sodus Bay composed Hall's Lower Green shale. In 1918 Chadwick[6] suggested that the basal shale at Niagara might be of Furnaceville age, and in 1919 Williams[7] likewise called it Furnaceville. Chadwick also stated that "it is surely not Maplewood." Further details of the history of the nomenclature may be found in Table I.

The formation is composed of smooth green shale, of similar megascopic appearance throughout. In addition to shale minerals it contains a small amount of quartz and some calcite.

[4] About 3 inches of green shale occur here between the Thorold and the Reynales, as subsequently noted.

[5] *Op. cit.*, pp. 59–60.

[6] G. H. Chadwick "Stratigraphy of the New York Clinton," *Geol. Soc. Amer. Bull. 29* (1918), p. 344.

[7] M. Y. Williams, "Silurian Geology and Faunas of the Ontario Peninsula, and Manitoulin and Adjacent Islands," *Canada Dept. Mines Geol. Surv. Mem. 111* (1919), pp. 47–148, Fig. 3.

TABLE I

NOMENCLATURE OF THE NIAGARA GORGE SECTION

Hall* 1843	Grabau† 1901	Kindle and Taylor‡ 1914	Chadwick§ 1918	Williams‖ 1919	Sanford
Niagara limestone	Guelph domomite Lockport limestone 1. Hydraulic cement beds 2. Crinoidal limestone 3. Geodiferous limestone	Gasport limestone — Lockport Dolomite		Eramosa beds Gasport dolomite Decew waterlime — Lockport Guelph — Niagara	
Niagara shale	Rochester shale	Rochester shale — Clinton Formation	Rochester shale	Shale — Rochester — Niagara	Rochester formation
Upper limestone Lower limestone Lower Green shale — Clinton	Clinton upper limestone Clinton lower limestone Clinton shale	Irondequoit limestone Wolcott limestone Sodus shale — Clinton Formation	Irondequoit limestone Reynales limestone Furnaceville(?) shale	Irondequoit dolomite Reynales (Wolcott) dolomite Furnaceville (Sodus) shale — Clinton — Niagara	Irondequoit formation Reynales formation Neagha formation
Greenish gray argillaceous or siliceous sandstone Red shale and sandstone Gray quartzose sandstone Red marly or shaly sandstone — Medina	Upper shales and sandstones Gray quartzose sandstone Red Medina shales — Medina Sandstones and Shales	Thorold sandstone Whirlpool sandstone — Queenston Shale Albion Sandstone	Thorold sandstone Medina sandstone	Thorold sandstone Red Lingula Gray Red — Grimsby Sandstone Cabot Head shale Manitoulin beds Whirlpool sandstone Queenston shale — Medina	Thorold formation

* James Hall, *Geology of New York, Part 4, Survey of the Fourth Geological District* (1843), pp. 32–117.

† A. W. Grabau, "Guide to the Geology and Paleontology of Niagara Falls and Vicinity," *Bull. Buffalo Society of Natural Sciences*, Vol. VII, No. 1, pp. 87–114.

‡ E. M. Kindle and F. B. Taylor, "Niagara Folio," *U.S. G.S. Folio 190* (1914), Fig. 5.

§ G. H. Chadwick, "Stratigraphy of the New York Clinton," *Geol. Soc. Amer. Bull. 29* (1918), p. 332, Fig. 1.

‖ M. Y. Williams, "The Silurian Geology and Faunas of Ontario Peninsula, and Manitoulin and Adjacent Islands," *Canada Dept. of Mines Geol. Surv. Mem. 111*, No. 91 (1919), Fig. 3.

The fauna is neither large nor plentiful. Grabau[8] notes: *Bythotrephis gracilis*, *Anoplotheca hemispherica* (*Coelospira hemispherica*), *Anoplotheca plicatula* (*Coelospira plicatula*), and possibly *Pterinea emacerata*. Williams[9] records a slightly longer list, including, in addition to some of the species mentioned by Grabau: *Rhynchotreta cuneata americana* (Hall), *Ctenodonta elliptica* (Hall), *Cuneamya alveata* Whitfield and Hovey, and *Orthoceras* 2 sp. He also definitely lists *Pterinea emacerata* (Conrad), which Grabau suggested.

The forms which I have observed are as follows: *Coelospira* sp., *Lingula*, n. sp., brachiopod, unidentified, *Pterinea emacerata* (Conrad), *Ctenodonta elliptica* (Hall), pelecypod, family *Modiolopsidae*, and cephalopods, 2 sp. It will be noted that *Coelospira hemispherica* (Sowerby) is omitted from this list. No specimens which could be placed in this species have come to my attention. Since the implications in the writings of Grabau and Williams are that it is one of the more common forms, it would seem, although it is true that the fossils are not plentiful, that, if present, it might be represented in the material studied by me.

Specimens from Grabau's collection at Columbia University, most kindly loaned by Dr. G. M. Kay, and labeled "*Anoplotheca hemispherica*," differ from that species, most notably in the presence of a well-developed medial fold and sinus, and are not referable to any species known to me. A tray containing a few incomplete specimens labeled "*Anoplotheca hemispherica* Sby., Clinton shale, Niagara Gorge, Grabau," was examined at the Buffalo Museum of Science. These specimens likewise differ from *Coelospira hemispherica*. *C. hemispherica* ranges down into the Reynales type of lithology to the eastward and might be represented in the Neahga, but, in my opinion, fragments of other forms have been mistaken for this species. The differences between *C. hemispherica* as it occurs in the pearly layers of the Sodus at Rochester and the material from the Niagara Gorge can be seen in Figures 1–4.

Most of the species found in the faunal lists are forms usually

[8] A. W. Grabau, "Guide to the Geology and Paleontology of Niagara Falls and Vicinity," *Bull. Buffalo Soc. Nat. Sci.*, Vol. I (1901), p. 96, and pp. 233–37.

[9] *Op. cit.*, p. 48.

associated with younger rocks, some forms ranging as high as the Rochester. Emphasis should also be placed on the fact that there are several species which have not as yet been found in sufficiently good condition for determination, or have not to my knowledge been found elsewhere, so that some of the faunal affinities remain unknown. From the paleontological point of view, the Neahga seems more closely related to the overlying formations.

Stratigraphically the Neahga does not, as has always been indicated, rest on the Thorold sandstone. Below it, and above the Thorold, is a bed of calcitic sandy rock containing brachiopods, the specimens of which are so poorly preserved as not to be specifically identifiable. It is entirely possible that there is a small unconformity between this bed and the Thorold, although a time break here would be rather difficult to prove. This possibility is suggested by the sudden change in lithology and the presence of the brachiopods mentioned above. These criteria and the stratigraphic relationships indicate that this calcitic bed is more closely related to the rocks above than to those below. It is not considered sufficiently important to warrant a special designation at this time.

I do not believe that at present any positive correlation of the lower Clinton shale is possible either on paleontologic or stratigraphic grounds, although the lower Clinton age of this shale seems certain. The suggestions of Chadwick and Williams concerning the Furnaceville age of the deposit may be correct, but this cannot be demonstrated by the use of a brachiopod, *Coelospira hemispherica* (Sowerby) which, it is my belief, does not occur in the formation. Moreover, should this fossil be one of the constituents of the fauna, its presence might be interpreted in several ways. *C. hemispherica* occurs abundantly above and not below *Hyattidina congesta*, which overlies the Neahga in the Niagara Gorge. This latter form occurs in the Reynales limestone at Niagara, and it will be shown that this formation either correlates with or is younger than the Furnaceville member of the Reynales at Rochester. It would seem, therefore, that the Neahga might correspond in age, at least approximately, to the Maplewood but that they were laid down either in separate basins or embayments, or under different environments, at least as far as conditions for life were concerned. The term "Neahga"

(meaning Niagara River), which can be used as a local name implying no correlation, is therefore proposed to designate this formation.

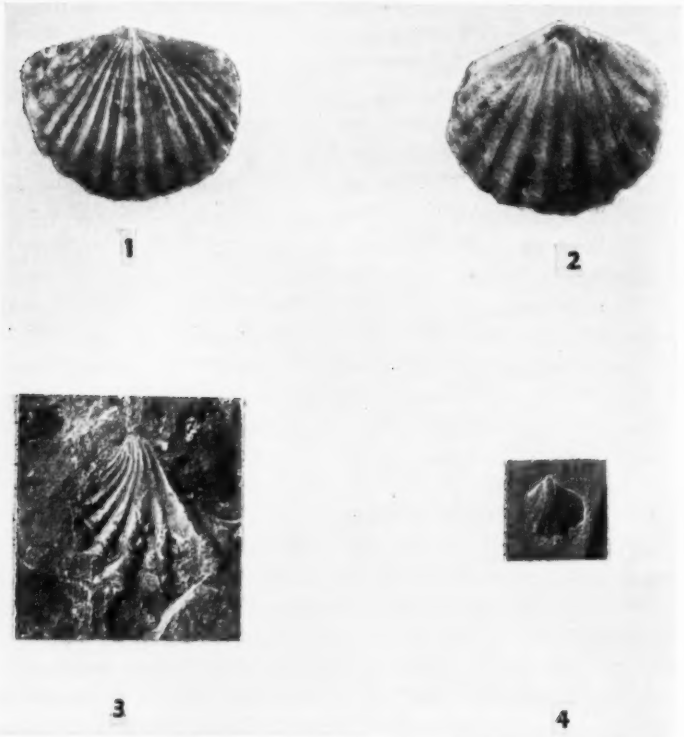

FIG. 1.—*Coelospira hemispherica* (Sowerby), dorsal valve, approximately ×3, Sodus formation, Genesee Gorge, Rochester, New York. Collected by J.T.S.

FIG. 2.—*Coelospira hemispherica* (Sowerby), ventral valve, approximately ×3, Sodus formation, Genesee Gorge, Rochester, New York. Collected by J.T.S.

FIGS. 3 and 4.—*Coelospira* sp., approximately ×3. Fig. 3, Neahga formation, Niagara Gorge, near Lewiston, New York. Collected by J.T.S. Fig. 4, Neahga formation, Niagara Gorge, near Lewiston, New York. Collection of Columbia University.

THE MAPLEWOOD

The Maplewood shale occurs only in the vicinity of Rochester, the type exposure being in the Genesee Gorge at Maplewood Park,

where it is 21 feet thick. It is a soft fissile green shale, in part of which there are great quantities of extremely thin, flat disks, varying somewhat in diameter, but generally smaller than a dime in this dimension. In spots they are so numerous that they mask the usual appearance of the rock, even though the lithology at that particular place does not seem to differ from that in those places where they are absent. These disks are calcareous. Their origin is still a mystery, no satisfactory theory having been brought forward to explain them. It is not definitely known whether they are of primary or secondary origin, or whether they are of organic or an inorganic nature. If they are not of organic origin, there are no fossils definitely known in the Maplewood formation.

At Glen Edith, on the east side of Irondequoit Bay, the lithology is similar to that of the Genesee Gorge, but the thickness has diminished to approximately 15 feet. Near Union Hill, a few miles farther east, a few inches of green shale below a hematite-rich bed may or may not correspond to this horizon. At points still farther east it is entirely missing.

The Maplewood shale was confused with the overlying purple shale (Table II), as will be subsequently explained, until 1918, when Chadwick[10] recognized that it was not of Sodus age and applied the present term.

Ulrich and Bassler[11] realized the true stratigraphic position of the Maplewood shale, but they believed that it extended farther to the east than it actually does in the outcrop. They interpreted the calcareous green shale in the sections at Wallington and Wolcott, described in the report of Newland and Hartnagel[12] as the Maplewood. There is a green formation at that horizon; but, as seen in the field in this vicinity, it is a sandstone (the Thorold) and not a calcareous shale. However, well cuttings supplied by Mr. Tracy Gillette from the McGuane well to the south indicate the presence of a green shale of the appearance of the Maplewood at this horizon, and this seems

[10] *Op. cit.*

[11] E. O. Ulrich and R. S. Bassler, "Paleozoic Ostracoda: Their Morphology, Classification, and Occurrence," *Maryland Geol. Surv., Silurian* (1923), p. 329.

[12] D. H. Newland and C. A. Hartnagel, "Iron Ores of the Clinton Formation in New York State," *N.Y. State Mus. Bull. 123* (1908), pp. 31–32.

TABLE II

NOMENCLATURE OF THE ROCHESTER SECTION

Hall* 1843		Hartnagel† 1906	Chadwick‡ 1917	Ulrich and Bassler§ 1923	Sanford 1930
Niagara limestone	Niagara Formation	Lockport dolomite			Lockport dolomite
Niagara shale		Rochester shale	Rochester shale	Rochester shale	Rochester Shale
5 Upper limestone of Clinton group	Clinton Formation	Irondequoit limestone	Irondequoit limestone	Irondequoit limestone	Irondequoit limestone
4 Second green shale		Williamson shale	Williamson shale	Williamson shale	Williamson shale
			Sodus shale	Zygobolba anticostiensis zone (Gun River) (Bear Creek?)	Sodus shale
3 Pentamerus limestone		Wolcott limestone	Reynales limestone	Reynales limestone	Reynales limestone
2 Oolitic or lenticular iron ore		Furnaceville ore	Furnaceville iron ore	Furnaceville iron ore	Furnaceville member
			Bear Creek shale		Brewer Dock member
1 Lower green shale		Sodus shale	Maplewood shale	Maplewood shale	Maplewood shale
Greenish gray argillaceous or siliceous sandstone. Included in the	Median Formation	"Gray band" (of the Medina)	(Thorold)	Thorold sandstone	Thorold sandstone
Medina sandstone		Upper Medina sandstone and shale Lower Medina shale	Medina sandstone and shale	Albion Queenston	Medina sandstone Queenston shale

* James Hall, *Geology of New York, Survey of the Fourth Geological District*, Part 4 (1843), pp. 32–117.

† C. A. Hartnagel, "Geologic Map of the Rochester and Ontario Beach Quadrangles," *N.Y. State Mus. Bull. 114, Paleontology 17* (1907), pp. 10–27.

‡ G. H. Chadwick, "Stratigraphy of the New York Clinton," *Geol. Soc. Amer. Bull. 29* (1918), pp. 341–53, Fig. 1.

§ E. O. Ulrich and R. S. Bassler, "Paleozoic Ostracoda: Their Morphology, Classification, and Occurrence," *Maryland Geol. Surv., Silurian* (1923) p. 347, Fig. 26.

to check the data collected by Newland and Hartnagel. Whether or not the green shale seen in the well is the Maplewood remains a question.

Similarity to the Neahga in stratigraphic position and perhaps in mode of origin has been suggested above; but in the absence of fossils and in the light of the limited east-west extent of the Maplewood along the line of outcrop, the latter term can be applied only to rocks occurring in the vicinity of Rochester.

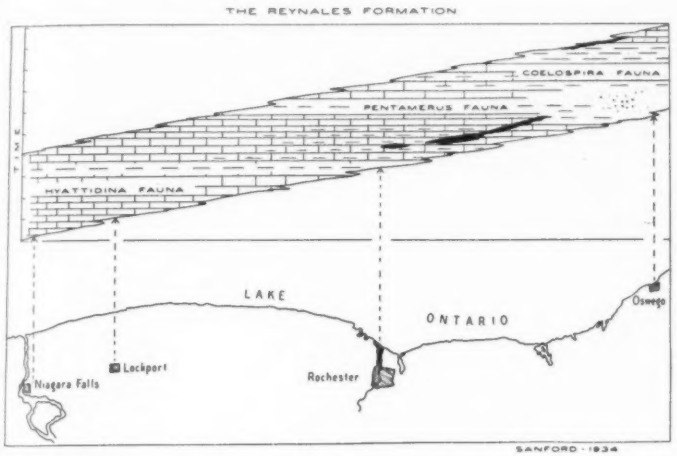

FIG. 5.—The Reynales formation showing migration across time and space, with lithologic and faunal changes.

REYNALES

The Reynales formation, confused with the Wolcott by the earlier workers, extends throughout the western part of the state as a recognizable unit but varies from west to east in lithology and faunal content. In the western sections it is a limestone; but progressing eastward it becomes more and more shaly, until at "Bear Creek" (or Black Creek)[13] it contains a large percentage of shale. In the vicinity of the Genesee Gorge and eastward the formation is characterized by hematite-bearing lentils, frequently termed "iron ores." At the Niagara River the formation rests on the Neahga shale and is

[13] Black Creek on the topographic maps.

§ G. H. Chadwick, "Stratigraphy of the New York Clinton," *Geol. Soc. Amer. Bull.* 29 (1918), pp. 341–53, Fig. 1.
§ E. O. Ulrich and R. S. Bassler, "Paleozoic Ostracoda: Their Morphology, Classification, and Occurrence," *Maryland Geol. Surv., Silurian* (1923) p. 327, Fig. 26.

overlain disconformably by the Irondequoit (Fig. 6). The Thorold sandstone is separated from the Reynales by only 3 inches of green shale at Lockport, and here the top of the Thorold is shaly and contains small black nodules. As at Niagara, the Irondequoit is the overlying formation. At Rochester the Reynales is divisible into three units, the lower two of which have been considered separate formations but which are more properly considered as members of the Reynales. Using the Reynales in this more comprehensive way,

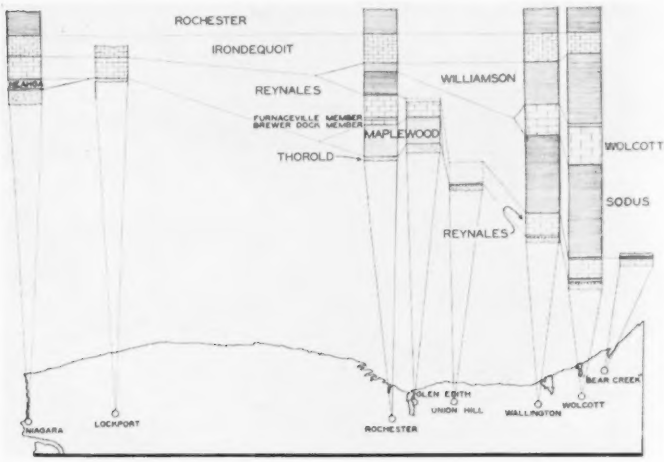

FIG. 6.—Correlation in the "Clinton" of Western New York

as was suggested by Ulrich and Bassler,[14] it is underlain by the Maplewood and overlain by the Sodus shale. East of Rochester, at Union Hill, it is underlain by the Thorold; the upper part is under cover; but at Ontario Center, as at Rochester, it is overlain by the Sodus. These conditions continue to the eastward in the area here discussed.

At Rochester the upper member of the Reynales contains a *Pentamerus* limestone, which was confused with the Wolcott limestone, a *Pentamerous* limestone of somewhat younger age occurring in the vicinity of Sodus Bay. Hence the term "Wolcott" was ap-

[14] *Op. cit.*, pp. 329–30.

plied to the Reynales at Rochester and Niagara until 1918, when Chadwick[15] demonstrated the true relationship of the two limestones and applied the present term to the formation, including in it, however, only the upper member at Rochester. The type locality named by Chadwick is at Reynales Basin, 8 miles east of Lockport, although as he says: "We must go to Lockport or the Rochester Gorge for the typical section."

It has been noted that the Reynales at Rochester is divisible into three units, the lower two of which have been considered separate formations but which are more properly considered as members of the Reynales. The lowest of these is the "Bear Creek" of Chadwick;[16] but inasmuch as the beds at Rochester, for which this term was intended, do not correlate with the type locality and as there is no need of a special term with reference to that locality, it is best to discontinue the use of the term. It is suggested, for the convenience of local workers, that the term "Brewer Dock member" be applied to that part of the Reynales which occurs below the hematite-rich bed at Rochester. The latter has been referred to as the Furnaceville; and although the bed at Rochester is not traceable into the one at Furnaceville as a lithologic unit the correlation is so close that the term is retained.

In the Niagara Gorge and at Lockport *Hyattidina congesta* characterizes the Reynales fauna. At Rochester this fossil is still present but is less conspicuous, being reported by Hartnagel[17] from the lower member. Ulrich and Bassler[18] report the *Hyattidina congesta* fauna also from the lower member. During the summer of 1932 it was collected from above the Furnaceville member and within a few inches of the shell limestone filled with *Pentamerus oblongus*, which is so conspicuous in the Genesee Gorge.

Proceeding eastward from Rochester, *Pentamerus* becomes less numerous, although still present at least as far east as Salmon Creek; but at this locality the fauna of the upper portion of the formation has a Sodus aspect. At "Bear Creek" the lower portion of the outcrop has a Reynales lithology, the shaly facies being prominent but not excluding the limestone, and with a hematite-rich bed at the top.

[15] *Op. cit.*, pp. 344–45.

[16] *Op. cit.*, pp. 342–43.

[17] Personal communication.

[18] *Op. cit.*, p. 327.

As has been pointed out by Ulrich and Bassler,[19] this outcrop contains a fauna which correlates with that of the purple shales so that paleontologically this rock is of Sodus age. The term "Bear Creek member" might be used to indicate this facies of the Reynales, which has a shaly Reynales lithology carrying a Sodus fauna. Near Martville, a few miles to the east of the area included in the present study, occur the sandstones and shales designated by Chadwick[20] as the Martville sandstone. These lie beneath strata considered by him to be of Reynales age, and overlie the Thorold. The stratigraphic position and lithologic character of these beds indicate that they are but a sandy, perhaps beach, phase of the Reynales. It is suggested, therefore, that the term "Martville," if it is to be retained, be used to designate a member of the Reynales. The Reynales represents a migrating sea; and although there is a lithologic change from west to east, this change is so gradual that the formation is recognizable throughout the area. This lithologic change is due not only to a migration across space but also to a migration across time, for, as has been demonstrated, the faunas to the west are older than the dominant fauna at Rochester, and the faunas to the east are still younger than those at Rochester (Fig. 5). These faunas overlap so that it does not seem probable that they can be explained as inhabitants of separate faunal provinces. However, this interpretation alone is not sufficient to explain the eastward increase in clastic materials; but when migrating physical conditions are considered as approaching an eastward or northeastward supply of clastic sediments, the idea seems plausible. Strictly speaking, the term "Reynales" denotes neither a time or a lithologic unit, but it does represent a stratigraphic unit, and a revision of the terminology to fit the details of the correlation would introduce more complexity than clarity.

SODUS

The disconformity between the Reynales limestone and the Irondequoit limestone at the Niagara River and at Lockport is occupied at Rochester by about 20 feet of shales. The lower 14 feet of these comprise the Sodus shale and are separated from the remainder, the Williamson, by a disconformity. They are predominantly purple in

[19] *Op. cit.*

[20] *Op. cit.*, p. 342.

color and contain thin lenses of shell limestones commonly known as "pearly layers." At the base are several feet of green shales which grade upward into the purple ones. These green shales characterize the base and have been noted elsewhere where this part of the formation can be seen. In the vicinity of Sodus Bay the formation has become much thicker, being approximately 50 feet thick, as shown by the drill cores described by Newland and Hartnagel.[21] In this vicinity it is separated from the Williamson by a *Pentamerus*-bearing formation, the Wolcott, composed of shale and limestone, at the top of which is a hematite-rich bed, the Wolcott furnace ore. This occurrence of *Pentamerus* above the Sodus caused the early workers to confuse it with the Maplewood at Rochester, which is also overlain by a *Pentamerus*-bearing limestone, the Reynales. In 1918 Chadwick[22] pointed out that the Sodus (of Sodus Bay) was younger than the Maplewood and corresponded in age to the purple shale at Rochester. More recently Ulrich and Bassler[23] suggested that the purple shale in the vicinity of Sodus is much younger, except for a few feet at the base, than the purple shale at Rochester. They state that the fauna of the purple shale at Rochester belongs to the *Zygobolba anticostiensis* zone of the island of Anticosti and that the fauna of the purple shale at Sodus belongs to the *Z. decora* zone of that island. On Anticosti there are about 600 feet of sediments between these two ostracod zones, and Ulrich and Bassler feel that between the time when the purple shales were laid down at Rochester and the time when they were laid down at Sodus "the process of marine sedimentation must have been interrupted and discontinued for a long time in New York."

Twenhofel[24] has since shown that *Z. decora* occurs above *Monograptus clintonensis* on the island of Anticosti. Inasmuch as the purple shales at both Rochester and Sodus occur below the *Monograptus clintonensis* zone, it hardly seems probable that either of them could contain the *Z. decora* fauna. An examination of the faunas of the purple shale from Rochester and Sodus Bay indicates that the *Z. anticostiensis* fauna is present at both localities, although the fossils from Sodus Bay are more numerous and varied.

[21] *Op. cit.*, p. 31. [22] *Op. cit.*, pp. 327–68. [23] *Op. cit.*, pp. 368–72.

[24] W. H. Twenhofel, "Geology of Anticosti Island," *Canada Dept. of Mines Geol. Surv. Mem. 154* (1928), "Geological Series," No. 135, pp. 36 and 96.

Two species of *Monograptus* found at the outcrop on Second Creek near Alton are of special interest. The following is a quotation from a letter written by Dr. Rudolf Ruedemann who very kindly identi-

TABLE III

NOMENCLATURE OF THE SECTION IN THE VICINITY OF SODUS AND WOLCOTT

Hall* 1843	Chadwick† 1917	Ulrich and Bassler‡ 1923	Sanford 1930
Niagara limestone			Lockport dolomite
Niagara shale		Rochester shale	Rochester shale
5 Upper limestone of Clinton group	Irondequoit limestone	Irondequoit limestone	Irondequoit limestone
4 Second green shale	Williamson shale	Williamson shale	Williamson shale
(Ore)	Wolcott furnace ore	Ore	Wolcott furnace ore
3 Pentamerus lime-	Wolcott limestone	Wolcott limestone	Wolcott limestone
1 Lower green shale	Sodus shale	Sodus shale (equal to Jupiter River) Zygobolba anticostiensis zone	Sodus shale
	Reynales limestone	Reynales limestone	Reynales limestone
	Furnaceville iron ore	Furnaceville iron ore	Furnaceville iron ore
	Bear Creek shale		
		Maplewood shale	
Greenish gray argillaceous or siliceous sandstone. Included in the		Thorold sandstone	Thorold sandstone
Medina sandstone		Albion	Medina sandstone

* James Hall, *Geology of New York, Part 4, Survey of the Fourth Geological District* (1843), pp. 32–117.

† G. H. Chadwick, "Stratigraphy of the New York Clinton," *Geol. Soc. Amer. Bull. 29* (1918), pp. 341–53, Fig. 1, p. 332.

‡ E. O. Ulrich and R. S. Bassler, "Paleozoic Ostracoda: Their Morphology, Classification, and Occurrence," *Maryland Geol. Surv., Silurian* (1923), p. 347, Fig. 26.

fied them: "They . . . . belong to the British species *Monograptus runcinatus* Lapworth and *M. becki*, two species that occur in a zone (22) that lies below the zone (23) where *M. priodon* (corresponding to our *M. clintonensis*) begins." This not only allows a closer corre-

lation with the British section but also makes it even more obvious that the *Z. decora* zone cannot appear in the Sodus shale.

It seems apparent that the purple shales at Rochester and Sodus Bay are of the same age, and may both be termed Sodus formation, although Sodus conditions of sedimentation were maintained longer at Sodus than at Rochester.

CONCLUSIONS

The conclusions herein presented are based on stratigraphic and faunal evidence. No attempt is made to discuss the sedimentation except in a general way.

The Neahga and the Maplewood occur at the same stratigraphic horizon and are of approximately the same age. It is not possible to make a thorough interpretation on the basis of the information obtained from the outcrops, but the accumulation of subsurface data may permit a more complete correlation. Obscure stratigraphic relationships are due to the erratic nature of the encroaching "Clinton" seas and the irregular condition of the surface over which they were advancing.

The Reynales is a complex unit. Its faunas, age, and the character of the sediments vary from place to place. The present interpretation calls attention to several factors:

1. A shoreline migrating from west to east in New York State.
2. Clastic sediments supplied most rapidly from the east and north.
3. The appearance of progressively younger faunas as the formation is followed from west to east.

Because of these factors the Reynales is considered to migrate across time as well as across space, although it still remains a formational unit.

The purple shales overlying the *Pentamerus* limestones at Rochester are shown to be of Sodus age.

The stratigraphic units discussed are designated as formations. It may appear that the Neahga and Maplewood are not of sufficient significance to be so classed; but if they are classed as members, the question arises as to just what they are members of. This usage follows the suggestions of the Committee on Stratigraphic Nomenclatures,[25] as each is a "fundamental unit in the local classification of rocks" and each is a "genetic unit."

[25] *Geol. Soc. Amer. Bull. 44* (1933) (Part 2), p. 430.

PHYSIOGRAPHIC HISTORY OF THE UPPER ARKANSAS RIVER VALLEY AND THE ROYAL GORGE, COLORADO

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Northwestern University

ABSTRACT

The physiographic history of the Arkansas River Valley between Leadville and Canon City is recorded in a descending series of benches, pediments, and alluvial plains flanking the present river. The highest of these seem to correspond to the Rocky Mountain peneplain which has been regarded as Miocene in age. Remnants of a later "broad-valley" stage lie above the flight of seven inner terraces developed in part on Pliocene sediments. Of these, two are preglacial in age, four correspond to glacial stages, and the lowest is postglacial.

The upper Arkansas River flows for a distance of some 120 miles through the Rocky Mountains of central Colorado, from Leadville to Canon City (Fig. 1). In the 60 miles above Salida the valley trends south-southeast between the Sawatch Range on the west and the Mosquito Range and Arkansas Hills on the east. South of Salida the Arkansas Valley is continued structurally by Wet Mountain Valley and Huerfano Park, but the river does not flow through these; instead, it cuts through the low mountains east of Salida and emerges from the Rockies at Canon City, thus dividing the Wet Mountains on the south from the dissected mountainous upland surrounding Pikes Peak on the north.

Throughout this distance the Arkansas Valley may be divided into sections (Fig. 1) which belong to two physiographic types: (1) broad basins characterized by steplike terraces on one or both sides of the river, and (2) narrow stretches of canyon cut in rock, with occasional terrace remnants yet preserved. The bordering mountain ranges lie well back from the axis of the Arkansas Valley, the narrow canyon sections having been cut in pediment-like uplands of a former "broad-valley" erosional stage. In the margins of the basins above Salida and below Leadville glaciers from the higher mountains deployed and built moranes associated with certain terraces marking stages of river development. In the canyon section between these two basins large glaciers pushed nearly or quite to the river channel, which thus is flanked by moraine on one side and bedrock on the other. The traveler cannot help being impressed by the contrast be-

tween these canyon sections, where the river is flanked by rugged rock walls or rough bouldery moraine, and the open basins in which

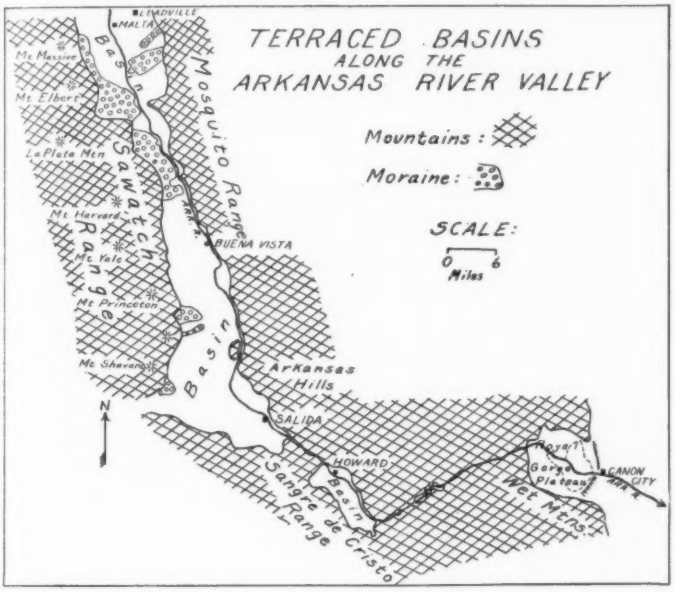

FIG. 1.—Map of the Arkansas River valley between Leadville and Canon City. The basins are occupied by Tertiary sediments and by boulder-covered terraces of Pliocene and Quaternary age.

broad terraces strewn with great boulders rise in a flight above an open inner valley.

EARLIER STUDIES

Certain features of, or pertaining to, the geomorphology of the upper Arkansas River Valley have been previously described by Davis,[1] Westgate,[2] Capps,[3] Campbell,[4] and Emmons, Irving, and

[1] W. M. Davis, "Glaciation of the Sawatch Range, Colorado," *Bull. Harvard Mus. Comp. Zool.*, Vol. XLIX (1904), pp. 392–404; "The Colorado Front Range," *Annals Assoc. Amer. Geog.*, Vol. I (1911), pp. 21–87.

[2] L. G. Westgate, "The Twin Lakes Glaciated Area, Colorado," *Jour. Geol.*, Vol. XIII (1905), pp. 285–312.

[3] S. R. Capps, "Pleistocene Geology of the Leadville Quadrangle, Colorado," *U.S. Geol. Surv. Bull. 386* (1909).

[4] M. R. Campbell, "Guidebook of the Western United States, the Denver and Rio Grande Western Route," *U.S. Geol. Surv. Bull. 707* (1922), pp. 90–121.

Loughlin.[5] A brief summary of the most significant facts brought out in these early studies is presented by Fenneman.[6] An incisive analysis of the geomorphic features of the region near Leadville has been given by Behre[7] and has served as the groundwork for the present study. The facts and conclusions presented herein are based on field studies pursued in the summers of 1932 and 1933 by the author as a member of the Northwestern Geological Field Party in Colorado. The kind assistance and criticism of Professors C. H. Behre, Jr., and J. T. Stark, of Northwestern University, and of Mr. T. S. Lovering, of the U.S. Geological Survey, are gratefully acknowledged.

BASINS AND TERRACES OF THE UPPER ARKANSAS VALLEY

The Arkansas Valley between Leadville and Canon City includes four broad basins developed in weak sedimentary formations, separated by stretches of narrow canyon where the river is intrenched in resistant rock. The first basin, in which Leadville and Malta are located, is about 15 miles long by 12 miles broad and extends to a point about 5 miles north of the station of Granite. Within this basin are high boulder plains or terraces described by Westgate, Capps, and Behre (Fig. 2), which rise outward at angles ranging from about $1\frac{1}{2}°$ near the Arkansas River to 4° or more near the bordering mountains, where they grade into rock shoulders or pediments thinly veneered with boulders. At the south, near Hayden, the terrace gravels play out; and these high terraces are developed not on bedrock but on deposits of white to buff arkosic sands and clays. Similar deposits penetrated beneath the terrace gravels by mine shafts in the vicinity of Leadville have been described by Loughlin under the name of "lake beds."[8] These arkosic clays and sands underlie terrace deposits believed to be preglacial and are probably of Tertiary age, an interpretation that is further supported by their similarity to Pli-

[5] S. F. Emmons, J. D. Irving, and G. F. Loughlin, "Geology and Ore Deposits of the Leadville Mining District, Colorado," *U.S. Geol. Surv. Prof. Paper 148* (1927), pp. 1–20.

[6] N. M. Fenneman, *Physiography of the Western United States* (New York: McGraw-Hill, 1931), pp. 92–132.

[7] C. H. Behre, Jr., "Physiographic History of the Upper Arkansas and Eagle Rivers, Colorado," *Jour. Geol.*, Vol. XLI (1933), pp. 785–814.

[8] Emmons, Irving, and Loughlin, *op. cit.*, pp. 17–20.

ocene deposits in the basins farther south and by the presence of mammalian remains in these beds where penetrated by a mine south of Leadville.[9] The terraces in the Leadville basin are classified as follows: (1) High boulder terrace remnants lying east of the river and between 300 and 400 feet above the latter at their riverward edge. These are Behre's preglacial, or No. 1, terrace. (2) High terrace remnants west of the river and about 220 feet above the latter at

FIG. 2.—View to east across Arkansas River valley toward seventh or highest terrace about 2 miles south of Leadville. Mosquito Range in distance.

their eastern edge. These constitute Behre's early glacial, or No. 2, terrace. (3) Three erosional terraces east of Hayden, developed on bedrock and on the Tertiary beds mentioned above at levels about 100, 250, and 450 feet above the present river. The 450-foot level seems to correspond to the preglacial boulder terrace (Behre's No. 1) and the 250-foot level to the early glacial terrace (Behre's No. 2). The lowest, or 100-foot, erosional terrace is of undetermined age, but appears to lie well above (4) a Wisconsin glacial terrace which forms much of the valley floor.

From the south end of the Leadville basin to a point about 8 miles

[9] *Ibid.*, p. 19.

below the station of Granite, the Arkansas River lies in a narrow canyon cut mainly in crystalline rock, though at some places the western canyon wall is rugged moraine dropped by glaciers from the Sawatch Mountains. This canyon opens into a basin about 37 miles long by 10 miles broad, containing the towns of Buena Vista and Salida. Here the river flows between boulder terraces that rise toward the bordering mountains like a flight of giant stair steps. The lower inner terraces are the youngest. Nearer the sides of the basin are

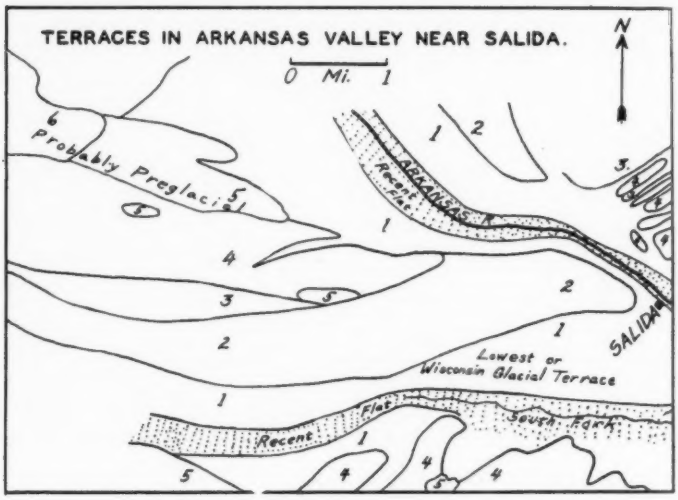

FIG. 3.—Distribution of terraces west of Salida, Colorado

older and higher terraces which owe their preservation to the fact that later stages in river development apparently were less complete than the early ones. About 5 miles west of Salida six terraces lie at levels respectively 15, 65, 135, 200, 265, and 350 feet above the recent flat of a tributary of the Arkansas River (Fig. 3). The first of these terraces probably corresponds to the Wisconsin glacial stage, and the highest two levels appear to be older than any of the known glacial stages in this region.

At many points the boulders of these terraces form merely a surface veneer over underlying arkosic and gypsiferous clays, sands, and

fine gravels that dip sharply toward the southwest. Mammalian remains discovered near Salida indicate the age of these beds to be Pliocene.[10] The several terrace levels now represented by widely scattered terrace remnants throughout this basin indicate successive periods of fan development and pediment cutting in which these features were graded to an early Arkansas River flowing between mountains less high than at present. That these pediments, fans, and terrace levels developed in late Pliocene and Pleistocene time is indicated by (1) their relation to Pliocene beds at Salida, and (2) a definite correlation of glacial moraines with the lower and interme-

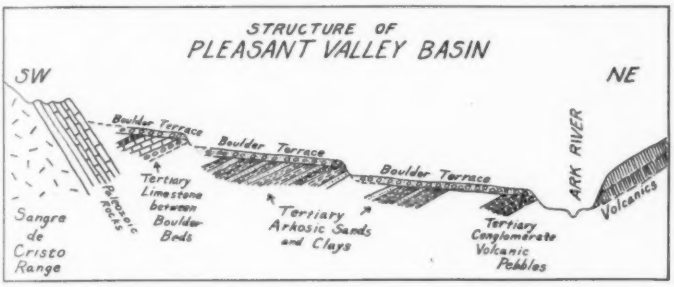

FIG. 4.—Geologic section 4 miles in length across Pleasant Valley near Howard. The southwestward-dipping clays, sands, conglomerates, and limestone are believed to be Pliocene in age.

diate terraces at several places, notably Chalk Creek Valley near Mount Princeton.

About 10 miles southeast of Salida the river enters Pleasant Valley, a basin approximately 11×4 miles in size, and similar in all respects to the Buena Vista–Salida basin. Terraces occur mainly on the west side of the Arkansas River at levels which, when projected to the center of the valley, lie approximately 30, 70, 150, 230, and 310 feet above the present river. These terraces have a veneer 10–50 feet thick of boulders, highly decayed in the oldest and highest levels, lying upon the beveled surface of arkosic clays, sands, and fine gravels similar to the Pliocene deposits observed upstream (Fig. 4) and, like them, dipping to the southwest. That this structure is fa-

[10] The remains are a molar of *Pliohippus leidyanus* and an astragalus of an unidentified camel. Identified by Dr. Alfred Romer.

vorable for landsliding is attested by moraine-like accumulations of débris in several of the ravines that trench the higher terraces.

The last intermontane basin through which the Arkansas River flows is the Royal Gorge plateau, an elevated tract 8 or 10 miles in diameter lying just west of Canon City. The plateau retains parts of

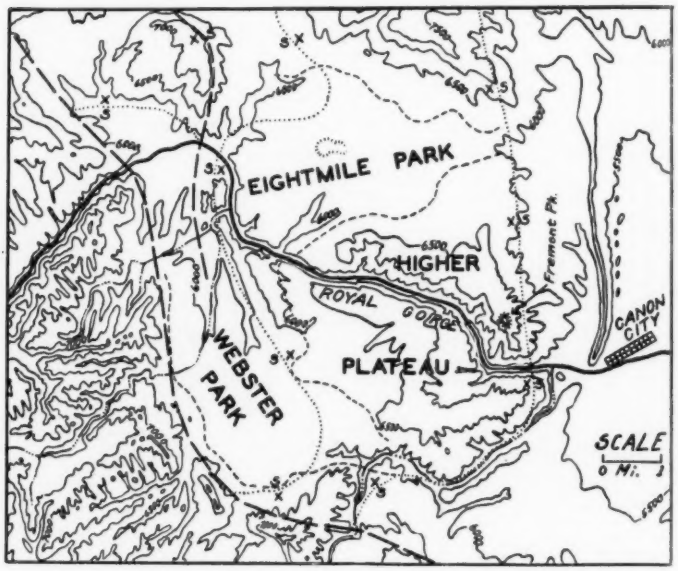

FIG. 5.—Map of region west of Canon City, showing higher plateau (6,800 feet) and lower plateau (6,100 feet) into which Royal Gorge is cut. Faults are shown as heavy dashed lines and boundaries between sedimentaries (*S*) and crystallines (*X*) as dotted lines.

two erosion surfaces: on the east, a somewhat rough upland about 6,800 feet in altitude, developed on crystalline rocks and containing a monadnock-like residual in Fremont Peak; and on the west, a lower surface with extensive flat areas about 6,100 feet in altitude, but in part dissected into broad valleys of a later erosion cycle (Fig. 5). This surface, known south of the Arkansas River as Webster Park and north of the river as Eightmile Park, is developed in large part on sedimentary formations of Mesozoic age. Both the 6,100-foot and

the 6,800-foot surfaces lie above the projected level of the highest alluvial terraces found upstream. However, such terraces are well developed above the Royal Gorge at Parkdale, where they occur at levels of approximately 50, 80, 130, 200, 310, and 400 feet above the present river; and below the Royal Gorge near Canon City, where their projected levels lie approximately 15, 80, 100, 140, and 200 feet above the river. Five miles below Canon City terraces occur at ap-

FIG. 6.—Well-cemented boulder bed of terrace resting on dipping Fountain formation 2 miles northwest of Canon City.

proximately 20, 60, 100, 130, 200, and 300 feet above the present river. The boulders of these terraces form at many places merely a thin cover over a rock surface developed by planation (Fig. 6). They indicate six or seven gradational stages in the development of the Arkansas River Valley that are practically unrecorded in the precipitous walls of the Royal Gorge.

Alluvial and erosional terrace remnants occur not only in the several basins of the upper Arkansas Valley but also in the narrower stretches of canyon. Although these fluvial terraces are poorly preserved, they lie close to the center of the valley and thus indicate accurately the levels of former stages in valley development. On the

contrary, the extensive basin terraces are not true fluvial terraces, but alluvial fans graded to cut rock pediments on the basin margins and to the former level of the Arkansas River in the middle of each basin. These basin terraces slope inward at angles ranging from 1° to 4° or more; whereas the general downstream descent of all terrace remnants of the same age is not very different from that of the present river, namely, 22–70 feet per mile or from one-fourth to three-fourths of 1°. Thus, if either the distance of terrace remnants from

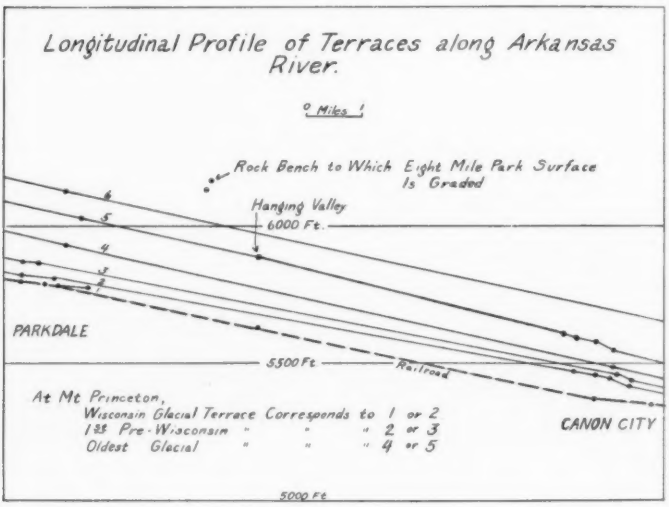

FIG. 7.—Profile of terrace levels developed along Arkansas River between Parkdale and Canon City.

the former position of the Arkansas River or the average slope of the terrace is unknown, the calculated level of the river at that stage of terrace development is likely to be considerably in error. In the present study the lower terraces were traced to the present river wherever possible, and the levels of the higher terraces determined by adding the successive intervals as determined elsewhere. When plotted on a longitudinal profile along the Arkansas River from Leadville to Canon City, the known terrace remnants fall into seven groups that lie about 20, 60, 80–100, 130–40, 200, 300, and 390

feet above the present river level. Six of these levels are shown in Figure 7.

OLDER PEDIMENTS AND ROCK BENCHES

The basins, terraces, and rock canyons of the upper Arkansas Valley are cut below an older upland level now broadly outlined by accordant rock ridges. Near the Sawatch, Mosquito, and Sangre de Cristo ranges these accordant ridges form foothills, and their character as dissected pediments is likely to be overlooked. East of

FIG. 8.—Wet Mountain Valley (in distance in center of photo) viewed from near Howard in Pleasant Valley. Wet Mountain Valley is part of a high pediment now partly dissected as shown in foreground.

Pleasant Valley they form the upland into which the Arkansas canyon is sunk. South of Pleasant Valley they form Wet Mountain Valley (Fig. 8), a broad intermontane basin deeply dissected only at its northern end. This valley lies between the Wet Mountains and the Sangre de Cristo Range and, together with Huerfano Park, is the structural continuation of the upper Arkansas Valley. However, the hypothesis that the Arkansas River formerly flowed southward through Wet Mountain Valley and Huerfano Park appears to be unsupported by physiographic evidence, despite their common alinement.

The oldest physiographic stage is recorded by certain shoulders or benches widely distributed throughout the Sawatch and Mosquito ranges at levels ranging from 11,500 to 12,500 feet (Fig. 9). These

FIG. 9.—High rock benches at about 12,000 feet near upper Taylor River, west side of Sawatch Range.

point to an ancient mature land that has been correlated with the Rocky Mountain peneplain of central Colorado by Behre.[11]

HISTORICAL RELATIONS OF THE GEOMORPHIC LEVELS

The 12,000-foot benches.—If Behre's correlation of the 12,000-foot benches of this region with the Rocky Mountain peneplain is correct, the belief of Van Tuyl, Lovering,[12] and others that this surface is Oligocene or early Miocene appears to be disproved by the general relations here. The southwestward dip of the Pliocene sediments in the Salida and Pleasant Valley basins requires that erosion surfaces developed prior to the deposition of these sediments be displaced vertically along the Arkansas Valley. Such vertical displacement is

[11] *Op. cit.*, p. 812.

[12] F. M. Van Tuyl and T. S. Lovering, "Contribution of the Cenozoic History of the Front Range, Colorado" (abst.), *Geol. Soc. Amer. Bull. 43* (1932), p. 170.

not present; on the contrary, shoulders at altitudes of 11,800–12,200 feet along the Continental Divide near the head of Quartz Creek, directly west of the dipping Pliocene sediments in the south end of the Salida basin, are identical in level with those observed in the northern Sawatch Range and in the Mosquito Range. Therefore the 12,000-foot benches appear to represent an erosion surface developed later than the deformation of the Pliocene sediments.

Pediments of the broad-valley stage.—True peneplanation did not occur at the present 12,000-foot level in the Sawatch and Mosquito ranges; but farther east the apparent superposition of certain streams on the Wet Mountains indicates base-leveling at this stage. Subsequent uplift led to the superposition of these streams and the development in late Pliocene time of the highest dissected rock pediments and the 6,800-foot plateau surface at the Royal Gorge, which probably correspond to the limited Mount Morrison surface elsewhere in the Front Range. Later Pliocene and Pleistocene time was marked by the terraces and basin surfaces whose history must be interpreted by their relations to the Tertiary sediments and the glacial moraines.

Relations of terraces to Pliocene sediments.—The occurrence in the Leadville, Salida, and Pleasant Valley basins of poorly consolidated, deformed sediments of Pliocene age has been hereinbefore described. These beds consist of arkosic sand and clay, gypsiferous in places, with occasional thin beds of fine quartzose gravel. Although interpreted by Emmons and others as "lake beds,"[13] their general character rather suggests deposition in broad playa basins in a region of low mountains and scant rainfall. In the Leadville basin these beds crop out only at the south, where they are generally horizontal except for local deformation associated with glacial drift and presumably due to ice shove. In the eastern part of the Leadville basin they underlie Behre's preglacial boulder terraces and are downfaulted against bedrock at the east. Mammalian remains found in one of the mines near Leadville are asserted by Loughlin[14] to indicate a late Pliocene or early Pleistocene age. Thus the high boulder terrace with its accompanying rock pediment is not older than late Pliocene; and because this terrace appears unrelated to glacial moraines and

[13] Emmons, Irving, and Loughlin, *op. cit.*, pp. 17–20. [14] *Ibid.*, p. 19.

their outwash, it is probably pre-Pleistocene. This terrace (No. 1 of Behre) corresponds to No. 5 or No. 6 of the seven terrace levels in the Arkansas Valley.

In the Salida and Pleasant Valley basins similar arkosic sands, clays, and gravels of proved Pliocene age dip southwestward, ranging from 10° to 35°. Near Salida these deposits exceed 500 feet in thickness and are beveled by the six boulder-capped terrace surfaces previously described. In the Pleasant Valley basin these sediments are

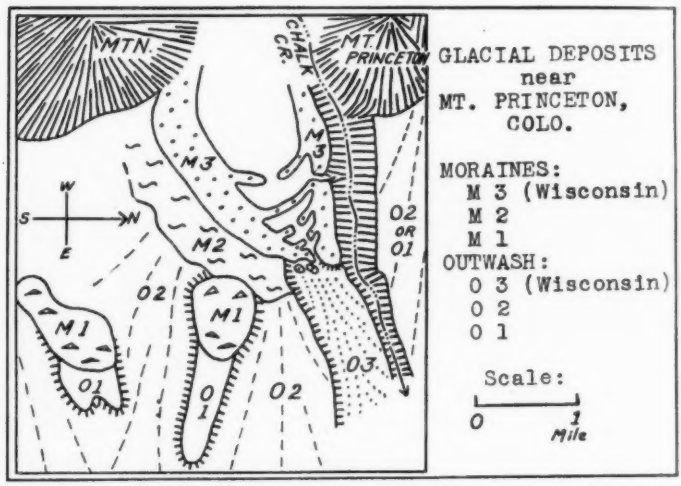

FIG. 10.—Moraines and outwash of three glacial stages near Mount Princeton, Colorado.

discontinuously exposed throughout the 4-mile width of the basin (Fig. 4). The lowest beds are at the northeast and consist of well-cemented gravels composed mainly of volcanic pebbles resembling the lava types found east of the Arkansas Valley, thus indicating that these lavas antedate the sediments. The sediments lying highest stratigraphically are at the southwest, close to the Sangre de Cristo Range, and consist of fresh-water limestone and beds of decayed boulders, dipping southwestward and beveled by the highest boulder-capped terrace. Thus, here also the development of basin and terraces occurred later than the deformation of deposits probably Pliocene in age.

Relations of terraces to moraines.—At several places it is possible to relate the Arkansas Valley terraces to glacial moraines. In Chalk Creek Valley near Mount Princeton are moraines and morainic remnants belonging to three glacial stages, each with its outwash plain or terrace (Fig. 10). The assignment of these to three distinct Pleistocene glacial stages is supported by their relative positions, degree of erosion, and degree of weathering of materials. There can be little doubt that the innermost and freshest morainic system belongs to

FIG. 11.—Postglacial boulder terrace (in foreground) and Wisconsin glacial terrace (in background at left) along Arkansas River, 8 miles north of Buena Vista.

the fourth, or Wisconsin, glacial stage. A projection of these glacial outwash levels to the Arkansas River indicates that the Wisconsin stage probably correlates with the second, or 60-foot, river terrace; the first pre-Wisconsin stage with the third, or 80–100-foot, terrace; and the earliest glacial stage with the fifth, or 200-foot, terrace.

The correlation of the second, or 60-foot, fluvial terrace with the Wisconsin glacial stage is further supported by their relationships near the upper end of the Buena Vista–Salida basin. Figure 11 shows the Arkansas River intrenched about 20 feet in a valley flat above which rises the bouldery Wisconsin moraine on the right (west) and a second terrace about 65 feet above river level on the left (east).

FIG. 12.—Large boulder in Wisconsin moraine, showing fluvial wear about its upper portion but not its lower. Near terraces of Figure 13.

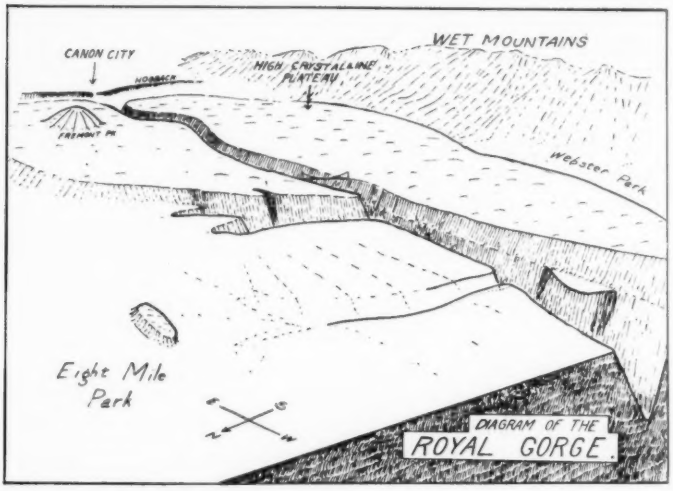

FIG. 13.—Physiographic diagram of the Royal Gorge region, showing the high crystalline plateau and the Eightmile Park physiographic surfaces.

Within the Wisconsin moraine there is evidence of stream abrasion at a level corresponding to the surface of the 65-foot terrace, some large boulders being stream-worn about their upper portions but not their lower (Fig. 12). Therefore the fluvial stage during which the 65-foot terrace was formed corresponds to the period when the Wisconsin moraine was built. The narrowness of the post-Wisconsin terrace here indicates that postglacial erosion has been relatively unimportant in the development of the upper Arkansas Valley.

In the Royal Gorge region the 6,800-foot and 6,100-foot surfaces point to two prolonged stages of stream erosion (Fig. 13). Both of these surfaces lie well above the highest of the six terrace levels in the Arkansas Valley and therefore are preglacial in age (Fig. 6). A ravine enters the Royal Gorge from Eightmile Park at a level approximately 200 feet below the Park surface and 270 feet above river level. This hanging valley coincides in level with the fifth terrace, probably of the earliest glacial stage.

SUMMARY OF PHYSIOGRAPHIC HISTORY

Shoulders or benches occurring at about 12,000 feet on the higher mountains appear to have been developed after the deformation of intermontane basin deposits of Pliocene age. These high benches mark a mature or post-mature surface possibly correlative with the Rocky Mountain peneplain of eastern Colorado. Later in the Pliocene an uplift of between 2,000 and 3,000 feet led to the development of a broad-valley stage now marked by extensive dissected rock pediments ranging from about 10,000 feet near Leadville to 6,800 feet at the Royal Gorge plateau. This stage may correspond to the Mount Morrison surface recognized farther north in the Front Range. In late Pliocene and Pleistocene time recurrent uplift has led to the development of seven well-marked sets of fluvial terraces to which are graded fans and pediments developed in the several basins. Their relations to moraines indicate that the lowest terrace level (about 20 feet above the present river) is postglacial; the second (60 feet) is the Wisconsin glacial terrace; the third (80–100 feet) is the first pre-Wisconsin glacial stage; the fourth (120–50 feet) is unrelated to known glacial moraines; the fifth (200 feet) is the earliest known glacial terrace here; and the sixth (300 feet) and seventh (390 feet) are older than the known glaciations. The narrow postglacial terrace has been trenched in the present stream cycle.

"CAROLINA BAYS" AND ELLIPTICAL LAKE BASINS

WILLIAM F. PROUTY
University of North Carolina

Early in 1933 a party[1] from the University of North Carolina began a geological and geophysical reconnaissance of portions of the Coastal Plain area of the Carolinas. Magnetometer readings were first made on a number of bays in the Myrtle Beach district, and a series of readings were run inland from Myrtle Beach to Florence and Darlington, South Carolina. In later surveys other cross sections of the Coastal Plain were made, and a number of bays and elliptical lake basins examined. These observations were made both in North Carolina and South Carolina and in both the inner and outer portions of the Coastal Plain. The magnetometer readings have given much valuable information concerning the depth, character, and structure of the crystalline rocks underlying the Coastal Plain.[2] The more steeply inclined Fall-zone peneplain of the buried crystallines has been shown to give place to a less steeply inclined buried peneplain farther inland; the buried Triassic basin near Florence, South Carolina, has been outlined and traced northeastwardly into North Carolina; other linear high magnetic areas of unusual character have been mapped.

In the earlier magnetometer investigation of the elliptical bays in the Myrtle Beach area the observations were, with few exceptions, largely made in, and not far outside, the bays. The differences in the readings were small and were thought at the time to be inconclusive. Later, in the neighborhood of Florence, South Carolina, and Fayetteville, North Carolina, anomalies of 250 γ were found in association with some of the bays (Figs. 1 and 2). On more recent trips into the lake country of southeastern North Carolina marked high magnetic

[1] Party composed of Dr. W. F. Prouty, Dr. G. R. MacCarthy, and J. A. Alexander, financed by a grant from the Rockefeller Fund for Research in Pure Science.

[2] G. R. MacCarthy, W. F. Prouty, and J. A. Alexander, *Jour. Elisha Mitchell Scientific Soc.*, Vol. XLIX (1933), No. 1, pp. 20–21.

areas were found in association with each of these elliptical depressions. That associated with White Lake has an anomaly of about

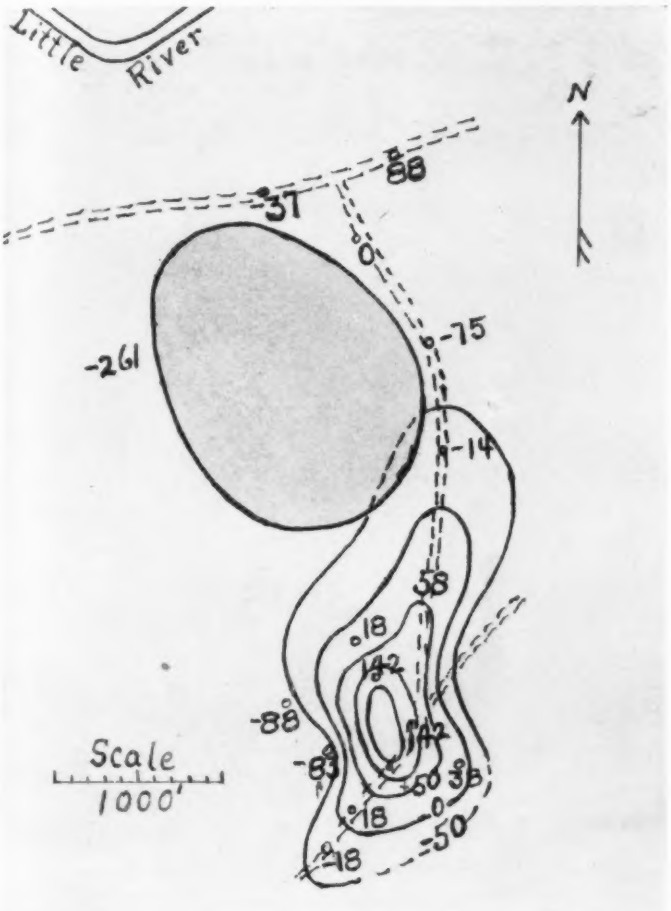

FIG. 1.—Shows high magnetic area southeast of the small bay near Little River in northwestern Cumberland County, North Carolina. Base map from aerial photo.

500 γ and that associated with a bay northwest of White Lake an anomaly of over 600 γ (Fig. 3). In fact, a positive anomaly has been

found in connection with most of the bays studied. Those in the Myrtle Beach area prove to be well defined though small.

Most of the work so far done with the magnetometer has been of a reconnaissance type; but the results seem to justify the conclusion that, in general, a magnetic high is to be found a little east of south

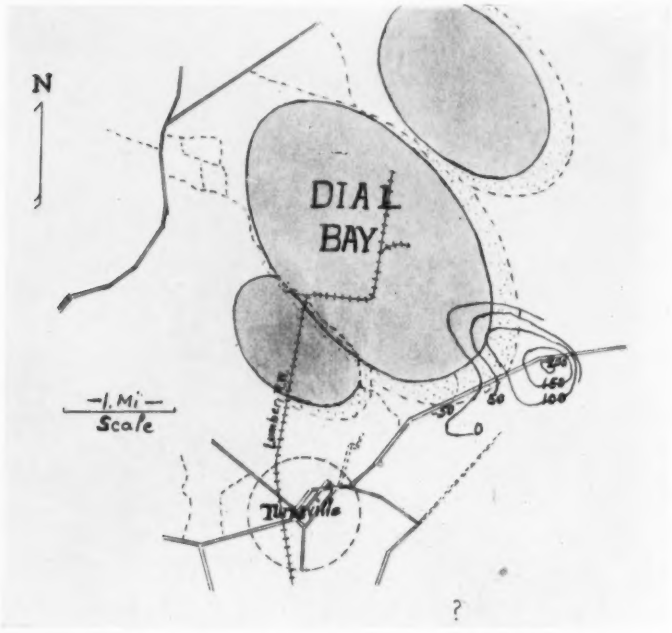

FIG. 2.—The high magnetic area located southeast of Dial Bay is now thought to belong to the smaller bay northeast of Dial Bay. If this is true, the magnetic high for Dial Bay should be about 1½ miles southeast of Turbeville. Base map after Mayesville Quadrangle and traverse. Locality 25 miles southwest of Florence, South Carolina.

of the southeast end of the elliptical bays and lake basins and at a distance about equal to the short diameter of the sand rim of such bay. The center of this high is at a greater distance from the bay than had been anticipated, which must mean, if the high is the result of a buried meteoric body (which seems reasonable), that the meteorite must either be buried at great depth or else it must have been deflected out of the original course by the more resistant crystalline

rocks beneath the Coastal Plain sediments. If we assume the meteoric origin of the bays and lake basins, the weaker anomalies to be

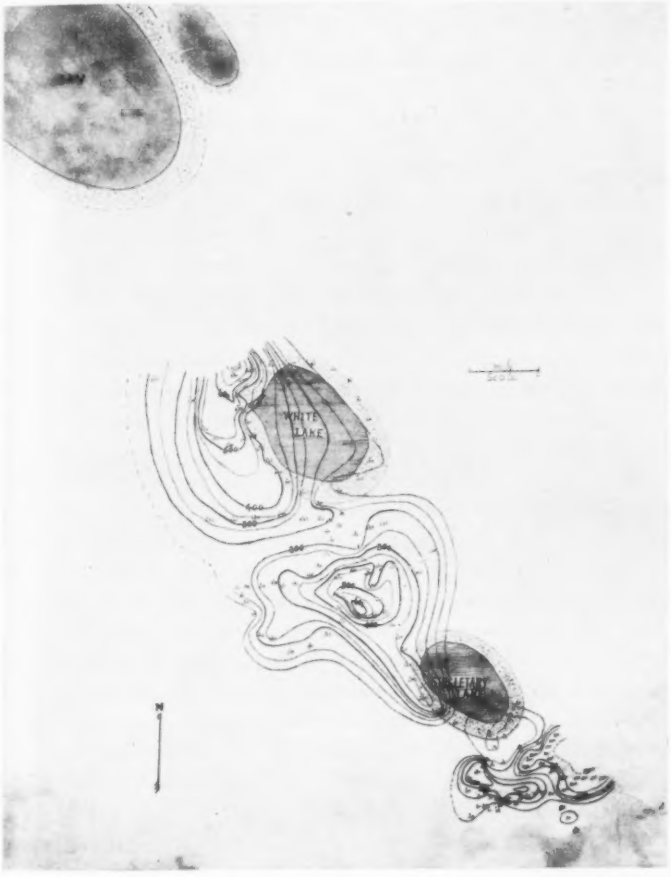

FIG. 3.—Magnetometer observations around Singletary Lake Basin, White Lake Basin, and large bay northwest from White Lake. The position of the center of the magnetic high in respect to the large bay and also in respect to White Lake conforms to the observed rule that this high is a little to the east of south of the southeast end of the elliptical depression and from it a distance about equal to the short diameter of the ellipse. Area to southeast of Singletary Lake is carried farther east by the influence of bays not shown on the map east of Singletary Lake. Base map after Soil Map, Bladen County, North Carolina.

found in the bays nearer the coast can be accounted for by the more rapid disintegration of the ferruginous bodies in the coastal area due to the presence there of sediments more highly charged with sodium chloride.

In general our observations support and supplement the views expressed by Drs. F. A. Melton and W. Schriever in the article on "Carolina Bays" published in the *Journal of Geology*[3] in 1933, and in Dr. Melton's reply to Dr. G. W. Cooke in a later article.[4] It is thought that the elliptical-shaped lake basins, largely in Bladen County, North Carolina, are of the same mode of origin as the elliptical bays; that many, or perhaps most, of these bays were originally lake basins; that many of them were, as is White Lake now, fed by lake-bottom springs. The chief effect of wind action on the rim sands seems to have been to partially fill in the elliptical basins from the southwest, and as a result of this most bays and lake basins are less convex on their southwest than on their northeast sides and the secondary or inner rims are covered on the southwest side of the bays (Fig. 4).

Our attention has been repeatedly directed to the presence of outer and inner rims in these bays and lake basins. In all cases where such double or triple rims exist, they tend to unite in a single rim toward the northwest end of the depression. In some cases the inner rim (ellipse) has an axis of little more than half that of the larger and outer rims. In such cases the two rims are more nearly equal in height than in those where the rims are close together at their southeast ends. It is apparent that these inner rims are formed by the forces responsible for the outer rims, since in all cases these inner rims are symmetrical to the axis of elongation of the larger outer rims.

It is difficult to visualize what would happen if a huge meteoric mass hit the earth with its necessarily high velocity, external high temperature, and its active air cushion. Would not sand be hurled to a great distance from the area of impact, and might not the air cushion and the steam pressure prevent the actual contact of sand and meteor near the surface, so that we might not expect to find much, if any, fused quartz? Would not steam explosions occur as the meteor

[3] *Jour. Geol.*, Vol. XLI (1933), pp. 52–66.

[4] *Ibid.*, Vol. XLII (1934), pp. 88–104.

sank deeper into the sediments? The secondary rims may best be explained in some such way.

A study of some of the soil survey maps reveals the presence of many of these double-rimmed bays. Perhaps one of the most conspicuous of these is to be seen on the Bladen County Soil Survey map toward the northwest ("Suggs Mill Pond" depression). The sand rim

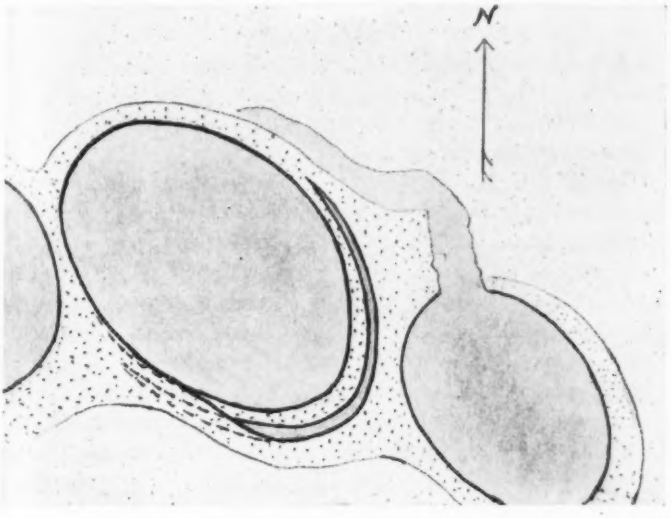

FIG. 4.—From aerial photograph, *New York Times*. Bay in Myrtle Beach area has well-developed secondary rim which is concealed on southwest side by the drift of sand from that direction.

is here indicated both by the soil survey symbols and by the position of the secondary roads which follow the less-timbered sandy rims.

Although the magnetometer observations point very strongly to the presence of buried meteoric bodies in association with these "Carolina bays" and elliptical lake basins, it is realized that more detailed and extensive mapping and checking of critical points is necessary before the meteoric theory can be firmly established.

A summary of our more important observations concerning the elliptical bays and lake basins in the Carolinas follows:

1. Most of these elliptical-shaped bays are found in southeastern

North Carolina, in South Carolina, and in the portion of the Georgia Coastal Plain near the Savannah River. Comparatively few are to be found in the Coastal Plain portion of northeastern North Carolina and of Virginia.

2. All elliptical-shaped bays and lake basins so far observed in this region have their long axes practically parallel. The axial direction is about northwest-southeast. It has no definite relation to the trend of the coast line of the region as a whole.

3. As a rule, the sand rim at the southeast end is higher and broader than elsewhere.

4. Sections of sand rims have been studied in a number of places, and absence of stratification noted.

5. In many cases two or more elliptical-shaped bays are found intersecting. In such cases the bay with the more perfect rim has, in its formation, completely destroyed the rims of the other bays in the area of overlap.

6. In all bays in which there is more than a single rim the additional rim is developed best at the southeast end. In general, this double or triple rim at the southeast end of the bay forms but a single rim at the northwest end of the bay. In some cases, the distance between the rims at the southeast end of the bay is considerable, and in such cases the heights of the rims are more nearly equal. A good example of this latter case is Suggs Mill Pond depression in the northwest corner of Bladen County, North Carolina.

7. The bays of the region have no regularity in regard to arrangement or topographic occurrence. A study of the available topographic and soil survey maps of the area seems to show that a larger proportion of the bays and elliptical lake basins are to be found on the higher interstream areas than in the stream lowlands.

8. The bays that are to be found along the inner edge of the Coastal Plain are as perfect and well preserved as are those along the Atlantic border. This seems to warrant the conclusion that the bays are all of about the same age.

9. Most of the elliptical-shaped lakes in southeastern North Carolina are rimmed, as are the typical bays, and in every way seem to correspond to the elliptical bays in mode of origin.

10. Some of these lakes, as for example White Lake in Bladen

County, have their chief water supply from lake-bottom springs. It is evident that many of the bays now largely filled with peat deposits were once lake basins.

11. Many of the bays, and particularly those nearer the coast, are unsymmetrical ellipses, their northeast sides being more convex than their southwest sides. A careful study of the conditions shows that this has been brought about by the partial filling of the bay lowland by drifting sand from the southwest.

12. Most of the elliptical bays and elliptical lake basins show a decided high magnetic area to the southeast of the bay. Measurements made with the magnetometer on a number of bays in all sections of the Coastal Plain seem to show about the same thing. In cases where the bays are fairly well separated and there is not too much interference of magnetic forces, there is usually a magnetic high area located a little east of south from the southeast end of the depression and at a distance about equal to the short axis of the bay.

13. Observations on bays near the coast give, in general, weaker anomalies than do those farther inland. Earlier observations made on the bays near the coast were less detailed and well placed than those taken later on bays away from the coast. Nevertheless, these earlier observations, when plotted, show a decided localized high area in the right spot to conform to the higher readings of the anomalies associated with the bays farther inland.

14. Some of the anomalies are comparatively high. That in connection with the White Lake Basin is about 500 γ, while the large bay to the northwest of White Lake shows an anomaly of about 600 γ. In comparison, observations on Meteor Crater, Arizona, have shown an anomaly of only about 45 γ.

TWO FOSSIL GRIZZLY BEARS FROM THE PLEISTOCENE OF OKLAHOMA

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ABSTRACT

Two fossil grizzly bears, here described from the Pleistocene of Oklahoma, are the only fossil grizzlies that have been reported from the Great Plains region. A brief survey of the literature is given, and the inadequacy of Merriam's classification is pointed out.

The two skulls described in this paper are unusual in their completeness and state of preservation. Their chief importance, however, lies in the fact that they extend the range of the grizzly bear into the Pleistocene of the Great Plains region, and are the only fossil grizzlies that have been described from this area.

In 1912, Hay[1] mentioned three species of the genus *Ursus* as having been found fossil in the United States: *Ursus procerus*, *U. vitabilis*, and *U. amplidens*. Of these, *U. vitabilis*, as described by Gidley,[2] is a small form represented by the lower jaws. It appears to have been about the size of *Euarctos americanus;* and there is no possibility, therefore, of the Oklahoma fossils belonging to this species. On the other hand, *Ursus procerus*, as figured and described by Hay,[3] shows many similarities to the fossils in question, and appears to be the only described American fossil bear of the grizzly type. This bear differs from the specimens here described in its more slender skull proportions and narrowness across the zygomatic arches. Also, the snout in *U. procerus* is not so deep, the frontofacial angle seems to be less, and the canines are not so strong as those of the skulls in question. Very little seems to be known about *U. amplidens*, although Cope[4] listed a posterior, lower molar which he stated is identical with

[1] O. P. Hay, "The Pleistocene Mammals of Iowa," *Iowa Geol. Sur.*, Vol. XXIII (1912), p. 475.

[2] J. W. Gidley, "Preliminary Report on a Recently Discovered Pleistocene Cave Deposit, Near Cumberland, Maryland," *Proc. U.S. Nat. Mus.*, Vol. XLVI (1914), p. 96.

[3] O. P. Hay, "The Pleistocene Period and Its Vertebrates," *13th Ann. Rept. Ind. Geol. Sur.* (1911), p. 772.

[4] Edwin D. Cope, "Synopsis of the Extinct Mammalia of the Cave Formations of the United States," *Proc. Amer. Phil. Soc.*, Vol. II (1869), p. 176.

Leidy's type specimen. The type specimen of *U. amplidens* as described by Leidy[5] from Natchez, Mississippi, contains two lower molar teeth. This fragment of the jaw is smaller than a corresponding portion of the jaw in *Euarctos americanus*, while the two teeth are equal in size to those of the grizzly bear. In 1923, Hay[6] listed two other members of this genus, one of which, *Ursus* (*Euarctos*) *floridanus*, is found both fossil and living in Florida; the other is an unidentified species. True brown or grizzly bears belonging to the genus *Ursus* are unknown in North America prior to the Pleistocene.

The first of these bear skulls to be described (University of Oklahoma Museum, No. 546) was found in a bed of river gravel of Pleistocene age, about 6 feet below the surface, near Lawton, Oklahoma, and was presented to the University by Mr. J. M. Stevens. The zygomatic arches posterior to the orbits are missing, as are also the lower jaws.

The facial part of the skull is relatively shorter than in *Ursus horribilis*. The frontal shield is short, broad, and flat, its posterior point being located 51 mm. back of the orbits. It is bounded posteriorly by well-defined temporal impressions which, at this end of the shield, mark the beginning of a pronounced sagittal crest 155 mm. in length. The nasal bones are solidly ankylosed and are 76 mm. in length. The lambdoidal crest strongly overhangs the occipital condyles. The paroccipitals and mastoids are well developed, the latter being inclined forward. The external auditory meatus is recurved and upturned. The postorbital processes are rather strongly developed. The postpalatal shelf is broad and flat, and is 55 mm. in width. The postpalatal notch is short and narrow. The palate is broad and deeply excavate. The basal length of the skull is 295 mm., the interorbital breadth 70 mm., and the palatal length 153 mm. The facial index, as calculated by dividing the interorbital breadth by the palatal length, is 0.46.

[5] Joseph Leidy, "Fossil Fragments Found in Association with the Remains of *Megalonyx*, in the Neighborhood of Natchez, Mississippi," *Proc. Acad. Nat. Sci. Phila.*, Vol. VI, No. 8 (1853), p. 303.

[6] O. P. Hay, "The Pleistocene of North America and its Vertebrated Animals from States East of the Mississippi River," *Car. Inst. Wash. Publ. No. 322* (February, 1923), pp. 350 and 382.

The teeth are badly worn, and the incisors and the first and third premolars are missing. The base of the canine is thicker and broader than in modern grizzlies. The fourth premolar is 13 mm. in length and 11 mm. in width; the first molar is 20 mm. in length and 15 mm. in width; the second molar is 30 mm. in length and 18 mm. in width (Fig. 1).

The second specimen (University of Oklahoma Museum, No. 601) was found about 4 feet below the surface in a river gravel of Pleistocene age, north of Cheyenne, Oklahoma, and was donated to the University by Mr. P. Chalmez. This fossil is somewhat more complete, the zygomatic arches being intact and the third left incisor still

FIG. 1.—Ventral view of skull of *Ursus horribilis oklahomaensis* from Lawton, Oklahoma. Length, 325 mm.

in place; otherwise its condition is similar to the one described from Lawton (Figs. 2, 3, and 4).

The zygomatic arches are widely expanded, being about equal to those of the modern grizzly, and well developed vertically. The palate is broad and excavate. The postpalatal notch is short and narrow, 25 mm. wide and 35 mm. deep. The postpalatal shelf is broad and flat, and 55 mm. in width. The paroccipitals and mastoids are well developed, the latter being inclined forward. The lambdoidal crest is strong and overhangs the occipital condyles. The temporal impressions, beginning at the postorbital processes, meet at a point 85 mm. back of the orbits, which is considerably farther back than in the specimen previously described, and may be due to sex difference, the longer temporal impression possibly being a female characteristic.

FIG. 2.—Profile view of skull of *Ursus horribilis oklahomaensis* from Cheyenne, Oklahoma. Length, 322 mm.

Figures 2, 3, and 4 are of the same individual.

FIG. 3.—Dorsal view of skull of *Ursus horribilis oklahomaensis* from Cheyenne, Oklahoma. Length, 322 mm.

FIG. 4.—Ventral view of skull of *Ursus horribilis oklahomaensis* from Cheyenne, Oklahoma. Length, 322 mm.; breadth, 202 mm.

The frontal shield is somewhat more convex over the sinus region of the skull in this specimen, but along the region of the suture between the frontal bones there is a tendency for the shield to be sulcate. The nasals are ankylosed and short, 78 mm. in length and 25 mm. in width. The frontonasal region is slightly elevated in the frontofacial plane. The lacrimal foramen is inside the orbit. The sagittal crest is 116 mm. in length. Basal length of the skull is 295 mm., zygomatic breadth 202 mm., palatal length 163 mm., interorbital breadth 71 mm. The cranial index, computed by dividing the zygomatic breadth by the basal length, is 0.68. The facial index, calculated by dividing the interorbital breadth by the palatal length, is 0.43.

In this specimen, also, the teeth are greatly worn, and the canines exhibit the same unusual degree of strength as in the skull previously described. The fourth premolar is 13 mm. in length and 11 mm. in width; first molar 20 mm. in length and 15 mm. in width, second molar 30 mm. in length and 17 mm. in width.

Comparisons.—During the Pleistocene epoch the climate of Oklahoma was at times considerably colder than at present, and arctic forms are occasionally encountered in the fauna. For this reason it was thought advisable to compare the two specimens with the skull of *Ursus maritimus*, as figured and described by Elliot.[7] The skull of *U. maritimus* is considerably longer and narrower in proportion than the two fossils in question. Likewise the angle between the facial region and the sagittal crest is less pronounced in *U. maritimus*.

As compared with three Canadian grizzly skulls, now in the University of Oklahoma collections, the fossils showed the following differences: the fossil skulls are shorter and narrower, the nasals shorter, the palate more concave, and the sagittal crest longer. Furthermore, the molars are shorter and narrower than are those of the recent grizzlies with which they were compared. Otherwise the skulls are very similar, and there is but little basis for the supposition that they constitute a separate and distinct species. For this reason Merriam's[8] discussion of the grizzly bears was carefully studied from a

[7] D. G. Elliot, "Synopsis of the Mammals of North America and the Adjacent Seas," *Field Columbian Mus. Zoöl. Ser.*, Vol. II, p. 310.

[8] C. Hart Merriam, "Review of the Grizzly and Big Brown Bears of North America," *North American Fauna, U.S. Dept. of Agri. Bur. of Biol. Sur. No. 41* (1928).

taxonomic standpoint, it being the desire of both the senior and junior authors to seek relationships rather than differences, and to avoid the creation of new species where the seemingly unnecessary burden of evidence points in a contrary direction.

The figures given by Merriam for the basal length, zygomatic breadth, palatal length, and interorbital breadth were used in the calculation of the facial and cranial indices, as explained above. These indices were calculated for the sixty-six species and varieties of grizzlies in Merriam's classification throughout his eight groups. The curves of these indices followed closely the normal distribution in both cases although slightly flattened, indicating a bimodality due, perhaps, to sex difference. The standard deviation for the cranial index was 8.18, the arithmetic mean being 67; and for the facial index, 6, the arithmetic mean in this case was 46. This is the type of curve to be expected in a closely related group of animals, and the figures given for the standard deviation are such as might be expected within a single species.

If the evidence and criteria are at hand for the splitting-up of the grizzlies into numerous species and varieties, it would seem, also, that the limits of variation within these species and varieties would be known and given as an important part of any workable classification. Merriam does not give the possible variation within his species, and does not seem to substantiate with sufficient evidence the validity of the groups and species that he has created. On the other hand, the facts presented in the preceding paragraph are, at least, slight evidence for the homogeneity of the grizzlies as a whole. Consequently, pending a more careful study, the grizzlies of North America are here all tentatively referred to *Ursus horribilis*, and two fossil skulls from Oklahoma are tentatively identified as *U. horribilis oklahomaensis*.

Thanks are extended to Dr. A. S. Romer for helpful criticisms in the preparation of this paper, and to Messrs. J. M. Stephens, of Lawton, Oklahoma, and Paul Chalmez, of Clinton, Oklahoma, for their donation of the foregoing described specimens to the University of Oklahoma.

DEVONIAN ICE IN PENNSYLVANIA[1]

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ABSTRACT

Evidence of the existence of Devonian ice is known from New York, the Maritime Provinces, and Alaska. Recent studies of the Devonian in Pennsylvania have revealed similar conditions there. It is believed that ice existed in early Portage time in Pennsylvania and that movements of this ice produced striations which have been preserved.

INTRODUCTION

Evidence of Devonian ice in New York State has been published by Clarke,[2] although he gave credit for its first recognition to James Hall in the latter's 1843 report on the Fourth District of New York (pp. 234–37). Clarke described striations in sandy shale of Portage age and pointed out their absence from contemporaneous, sand-free shale, due, he believed, to the incompetency of mud to retain the impressions. He concluded that they were made by tide-propelled ice scraping bottom. Although no glacial pebbles or morainal material were reported from New York, Clarke[3] noted the presence of closely contemporaneous, probably fluvioglacial conglomerate in Quebec. Evidence of Devonian ice has also been discovered in Alaska.[4] Schuchert,[5] reviewing Clarke's observations, analyzed associated faunas and concluded that the disappearance of corals and the reduction in numbers and the geographic relations of the associated cephalopods and pteropods indicate cooler water in late, than in middle, Devonian time.

In Pennsylvania, rocks of Portage age carry striae assignable to ice action. The contemporaneous faunas agree quite closely with

[1] Published with the permission of the State Geologist of Pennsylvania.

[2] J. M. Clarke, "Strand and Undertow Markings of Upper Devonian Time as Indications of the Prevailing Climate," *N.Y. State Mus. Bull. 196* (1917), pp. 199–210.

[3] J. M. Clarke, "Devonian Glaciation," *Geol. Soc. Amer. Bull. 26* (1915), abst., pp. 60–61.

[4] E. Kirk, "Paleozoic Glaciation in Southeastern Alaska," *Amer. Jour. Sci.*, 4th ser., Vol. XLVI (1918), pp. 511–15.

[5] C. Schuchert, "Winters in the Upper Devonian of New York and Acadia," *ibid.*, 5th ser., Vol. XIII (1927), pp. 123–32; "Winters in the Upper Devonian of New York. A Correction," *ibid.*, 5th ser., Vol. XIV (1927).

Schuchert's data, although corals do appear sparingly even into the Chemung in Pennsylvania. No floral evidence is available.

OCCURRENCE

During the 1933 field season the author's studies of the Devonian of Pennsylvania took him into east-central Huntingdon County. Among a number of stratigraphic sections examined, one is that exposed along the Trough Creek Branch of Tuscarora Creek and State Highway 829, which runs parallel to the branch between Blairs Mills and Shade Valley Post Office (Richdale). The section crosses at right angles a broad, shallow, symmetrical syncline of Devonian rocks. Helderberg limestone outcrops on the limbs $2\frac{1}{2}$ miles apart; the highest strata at the axis are Catskill continental red beds whose base intergrades with late marine Portage ("Ithaca") sandstone. Most of the eastern limb of the syncline is concealed; but west of the axis and below the "Ithaca" sandstone near a road fork $\frac{1}{2}$ mile east of Shade Valley Post Office, there outcrops 195 feet of platy to flaggy, gray or olive gray sandstone and sandy shale in the midst of which is a heavy sandstone. The trail *Pteridichnites biseriatus* Clarke and Swartz is found in the shalier beds. This fossil in south-central Pennsylvania appears to be almost confined to the otherwise nearly barren upper Portage shale (Braillier), which in eastern Huntingdon County are of early Ithaca age.[6] Its presence is important, as it fixes the age of the enclosing beds very precisely. Besides the trails, small, indeterminate, carbonized plant fragments are common, with mud chips and doubtful *Fucoides graphica* more rarely present.

A concealed interval of 20 feet separates this member from the next lower. This is a succession of 98 feet of gray or olive gray, platy to shaly sandstone which includes a *striated surface* in its upper part. The base of these beds is hidden in a concealed interval 125 feet thick, below which 10 feet of dark gray Burkett ("Genesee") shale rests upon the top of the Moscow formation (uppermost Hamilton) with its distinctive *Vitulina* zone. Fossils are probably absent from the striae-bearing member in this section; but in adjacent regions immediately subjacent beds carry a Naples fauna, chiefly small

[6] Bradford Willard, "Portage Group in Pennsylvania," *Geol. Soc. Amer. Proc. 1934 Meeting*, abst., p. 49.

pelecypods and pteropods with an occasional goniatite. These are correlated with the Harrell shale of more westerly sections.[7]

The striae are exposed across a smooth bedding surface about 30 feet square which was cleared during road construction. The strata strike N. 37° E. and dip 26° S. and the scratches trend N. 73° W. The largest scratches are grooves 30 mm. wide by 12 mm. deep, but the

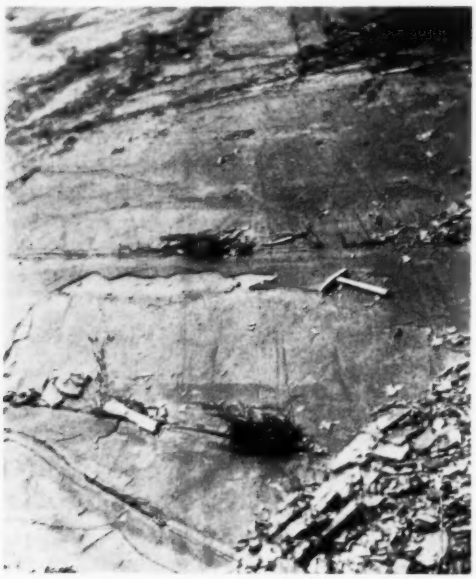

FIG. 1.—View of striated bed of lower Portage shaly sandstone in east-central Huntingdon County, Pennsylvania. Note that the striated surface passes beneath overlying strata in the background.

average is 10 mm. wide and 5 mm. deep. Some of them grow slightly wider toward their southern extremities, and some of the wider ones have fine striae within them (Fig. 2). The scratches are mutually parallel, suggesting a single movement of the body which produced them. The marks are so similar to Pleistocene glacial striae that, had they been found north of the terminal moraine instead of some 35 miles southwest of the nearest known signs of Pleistocene ice action,

[7] C. Butts, "Geologic Section of Blair and Huntingdon Counties, Central Pennsylvania," *Amer. Jour. Sci.*, 4th ser., Vol. XLVI (1918), pp. 523–37.

they might have been passed over without observing their true character. However, the age of the striations is absolutely established, since they are intraformational (Fig. 1, top portion) in the Portage.

FIG. 2.—Detail showing fine striae in larger grooves

INTERPRETATION

The age of the grooves eliminates the possibilities of Pleistocene ice action as a means of origin. Nor could they have been produced by human agency. Although exposed by stripping during road-building, there are few, if any, pick or steam-shovel marks. The extension of the grooves for 20–30 feet alone precludes such an origin. They bear no sign of organic remains such as bark impressions, carbonaceous residues, segmentation and the like—structures which might be expected were they imprints of vegetation. Furthermore, they are unbranched and all parallel to one another. The lack of branching, their straightness and parallelism, rule out their classification as rill marks. There is no particularly close resemblance to slickensides, no polishing on the freshest rock faces. No movement along this particular bedding surface can be detected; nor are the striae parallel to the dip, that is, perpendicular to the synclinal axis. Roots or other objects dragged over the surface are inadequate as an agent to explain these regular, symmetrical, straight lines. That they were formed by ice seems certain. That the ice was moved by water appears probable. Some grooves show what are taken to be faint chatter marks, which alone are strong presumptive evidence of ice action.

Clarke's shore-ice explanation is acceptable with modifications. By his theory blocks of ice are supposed to have been moved by the tides and currents up and down the strand either on the exposed beach itself through shoving above high-water mark or by dragging over a shallow bottom. The fact that the grooves widen is attributed not to the usual cause of wearing-down of points as in glaciation, for the materials of the sea bottom or beach would hardly have been capable of producing such a result, but to the advancing ice digging gradually deeper into the substratum. The degree of consolidation of the sediments must have been comparatively low, although admittedly high enough to record chatter marks. Clarke believed that the grooves he described were formed in unconsolidated material, an inference based upon their absence from shale which was too soft to hold the impressions. Lack of sharpness of the edges of some of the larger scratches, even though newly exposed and unweathered, is in partial agreement with such a theory, but the presence of the chatter marks already noted and the relative shallowness of the grooves are against such a supposition. The writer originally concurred with Clarke's belief that the striae were formed in entirely unconsolidated sediments, but discussion of this point with Professor C. O. Dunbar[8] has convinced him that a certain degree of consolidation was essential before grooving with chatter marks would be possible. Unconsolidated sand would probably be ploughed deeply by grounding ice; whereas partly consolidated sand, particularly sand containing a fair proportion of clay binder, even though not completely indurated, should resist ice-grooving much better. The presence of a disconformity at the striated surface, essential if the marks were made in hard rock subsequently buried, cannot be established, though there is no reason for not assuming a disconformable relation between the grooved bed and its capping.

The simplicity of the scratches points to a single movement of the ice. The marks must have been made during one continuous push, and not by a shuttle movement of tide-borne ice, because the lines are straight, parallel, and form a single, rather than a compound, pattern. It is thought that, rather than being necessarily of tidal origin, the movement may have been imparted to the ice by a current which carried it across a shallow bottom of partly consolidated,

[8] Letter of May 10, 1934.

clayey sand which it scratched in passing. It should be observed, however, that because of the simplicity of the striae, Professor R. T. Chamberlin[9] has suggested to the author the possibility of their having been caused by ice-rampart shoving, perhaps in a restricted area as a lagoon. This indeed comes nearer true glacial action than the already postulated cause through water-borne ice. Possibly, too, ice-rampart shoving may better explain the occurrence of chatter marks. Evidently, the precise process which produced the striations may yet remain to be determined.

CONDITIONS OF ORIGIN, ENVIRONMENT

The movement of ice was presumably near shore, judging from the varying character of associated sediments and from the fact that the marks are stratigraphically but a short distance below red beds, here probably of continental origin. It is furthermore known that the adjacent shore line was building rapidly south or southwest at this time.[10] A cool climate is admissible on faunal evidence, but fragments of carbonized plant material, undoubtedly of terrestrial origin, argue against an excessively severe climate. The presence of the ice may simply signify cold winters—a lowered mean annual temperature associated with late Devonian epeirogenic movements and orogeny. No evidence of widespread or severe ice action or continental glaciation at this time has yet been found in Pennsylvania.

CONCLUSIONS

Phenomena observed in sandy shale of early Portage age in south-central Pennsylvania are interpreted as indicating that ice was present near shore in the shallow sea off the west coast of Appalachia at that time. Paleontologic and stratigraphic data support this view. Evidently, Devonian ice occurred in Pennsylvania, as well as more or less contemporaneously in New York and Quebec. Schuchert has already advanced the proof of a cool late Devonian climate in New York and the Maritime Provinces. The area where such a climate obtained may now be extended southward to include Pennsylvania.

[9] Letter of August 25, 1934.

[10] Bradford Willard, "Early Chemung Shore Line in Pennsylvania," *Geol. Soc. Amer. Bull.* 45 (1934), pp. 897–908.

REVIEWS

The Jurassic System in Great Britain. By W. J. ARKELL. Oxford: Oxford University Press, 1933. Pp. viii+681; figs. 97, pls. 41. $7.75.

The object of the author, as expressed in the Preface, "to provide a general description of the Jurassic rocks of the British Isles," has been accomplished in a very successful manner. The subject is treated in eighteen chapters, grouped into four parts: Part I, "Classification and Chronology," Part II, "Distribution and Tectonics"; Part III, "Stratigraphy"; and Part IV, "Paleogeography."

Chapter i, which also forms Part I, "The Classification of the English Jurassic Rocks and the Partition of Jurassic Time," contains a comparison of the earliest tables of strata as developed by the pioneer stratigrapher, William Smith, together with tables made by Townshend and Buckland. The author follows very closely the classification of Buckland (1818) for the major subdivisions of the system. The importance of fossil horizons in Jurassic stratigraphy is emphasized; and the development and meaning of such terms as "secule," "faunizone," "biozone," "epibole," etc., form an interesting part of this chapter.

Part II, divided into two chapters, describes the troughs in which the Jurassic sediments were deposited and the diastrophic movements which occurred contemporaneous with deposition, together with their effects on sedimentation.

The stratigraphy of the Jurassic is covered in the fourteen chapters of Part III. Each chapter describes a major subdivision of the system, and all follow a uniform outline. The order of describing each formation is from south to north—from Dorset to Yorkshire and Scotland, and the Weald of Kent. The reading-matter is supplemented by numerous sketch maps, cross sections, columnar sections, and photographs of well-exposed outcrops. The stratigraphy is much clarified by the insertion of a table at the beginning of each chapter giving the stages, zones, formation names, and in some cases correlations between formations of two or more areas.

Although most of the book is taken up by the stratigraphic section, Part IV, containing less than 100 pages, forms an important part. Chapter xviii, on "Paleogeographical Conclusions," discusses the distribution

and environment of the more important marine organisms and the probable extent of the Jurassic sea over regions where no trace of sediments is to be found. This part also contains 13 plates of zonal index fossils; Appendix I, corals, reefs, geosynclines, and peneplains; and Appendix II, list of stages, containing 120 stage names which have been proposed for subdivisions of the system; Bibliography; and Index.

Both students and workers interested in the Jurassic system will find this well-organized and well-written book of great value, since it goes a long way in co-ordinating the work done in England, which is admittedly "the type of that of North-West Europe, in the most complete development anywhere yet known."

C. L. COOPER

"Fluid Mechanics of Salt Domes," by L. L. NETTLETON, in *Bulletin of the American Association of Petroleum Geologists*, Vol. XVIII, No. 9. Pp. 1175–1204.

An experimental study of the formation of salt domes was made on the basic assumptions (*a*) that the prime motive force is the density difference between the salt and the sediments; and (*b*) that both salt and sediments behave as highly viscous liquids. Some experiments were made with paraffin under mercury, but more of them with asphaltic petroleum under karo corn syrup. The doming movement had to be initiated by almost imperceptibly faint warping of the upper surface of the petroleum layer. There was a considerable range in the types of domes formed, depending on the relative viscosity of the two layers and the accidents of the mutual deformation. Many of the experimental domes closely simulated the forms of known salt domes. The most important new development of the experiments was the peripheral sink, the formation of which limited the flow of material to the core and led to the formation of rim synclines. A theoretical statement of the fluid mechanics is given.

This is one of the most important papers on the formation of salt domes, these being the first artificial salt domes which look like natural salt domes. One beauty of the method is that the whole growth of the dome is visible to the observers. The reviewer cordially recommends this paper to all students of salt domes, the method to geologists experimenting with the formation of geologic structures (for there well may be an element of fluid mechanics in many other types of structures), and the idea of fluid mechanics to the students of the structure of metamorphic-plutonic igneous rock terranes.

DONALD C. BARTON

Les ressources minèrales de la France d'outre-Mer. Tome II. Publications du Bureau d'Études Géologiques et Minières Coloniales. Paris, 1934. Pp. 436; figs. 59. Fr. 36.

The first volume of this series, upon coal, was issued at the end of 1933; the present volume treats of iron, manganese, chromium, nickel, tin, tungsten, graphite, glucinium, molybdenum, cobalt, titanium, and vanadium. The chapters, which were originally given as lectures at the Museum d'Histoire Naturelle in Paris, are by eight different authors. They bring together in concise form the geologic occurrence and composition of the ores, their annual production in various regions, and other useful information about these resources. These volumes are, in effect, a continuation of the publication *La géologie et les mines de la France d'outre-Mer* which appeared in 1932.

R. T. C.

Carte géotechnique de la Suisse. Feuille No. 1, Neuchâtel-Bern-Basel. With explanatory text in German and French by P. NIGGLI and F. DE QUERVAIN. Bern: Geographischer Kartenverlag Kümmerly und Frey, 1934. Pp. 65.

This map, on a scale of 1:200,000, features the nature of the surface materials and their utilization for the northwest quarter of Switzerland.

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This model, on a horizontal scale of 4 inches to the mile, portrays the western portion of the Ardnamurchan peninsula of Scotland and depicts the remarkable Tertiary ring intrusions of the district. The title suggests a piece of work chiefly of local interest; on the contrary, the descriptions, maps, and sections are of much interest to any structural geologist. Not only is the general Tertiary dike system of western Scotland very striking, but the development of ring dikes and cone sheets is perhaps nowhere better displayed than in this region.

R. T. C.

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